MESOSCALE AND SYNOPTIC PROCESSES IN MONSOONS

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I. **Mesoscale Processes in Monsoons** *(by Richard H. Johnson)*

1. Convection

A map of the six-year, TRMM global precipitation climatology is shown in Fig. 1. Some of the heaviest rainfall occurs in the regions of the Asian-Australian monsoon and the Indonesian maritime continent.

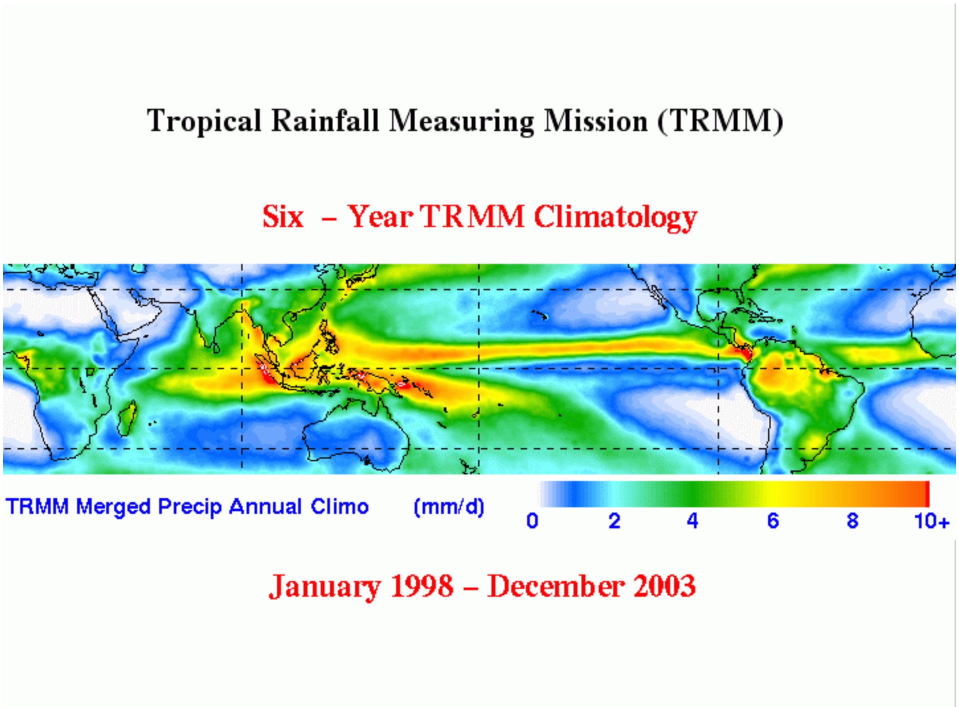

Figure 1. Six-year TRMM merged precipitation annual climatology.

The largest annual totals can be seen to occur in proximity to coastlines, suggesting possible roles of sea/land breezes and topographic effects in the precipitation mechanisms.

Knowledge of the nature of convective systems in the monsoon regions of the world, over both land

and ocean, is based mostly on case studies and limited field campaigns. There is evidence from these studies, however, to indicate that the characteristics of deep convective systems within the various monsoon regions bear a strong resemblance to each (Johnson and Houze 1987). In particular, convection tends to organize on the mesoscale and undergo an evolution characterized by a dominance of convective precipitation early in the life cycle followed by an upscale growth and development of stratiform precipitation on a time scale ~2-4 h and longer. The net result is a mesoscale convective system or MCS, defined by Houze (1993) as *a cumulonimbus cloud system that produces a contiguous area ~100 km or more in at least one direction*.

A recent study of MCS evolution for nearly 100 cases over the central United States has revealed three main patterns of organization (Fig. 2, from Parker and Johnson 2000). The three modes are convective lines with trailing (TS), leading (LS), and parallel (PS) stratiform precipitation. TS systems were the most common, accounting for ~60% of the cases, with the LS and PS each accounting for about 20%. TS systems have received considerable attention (e.g., Houze *et al.* 1990), but the occurrence of LS and PS systems is not insignificant, and there is evidence they are important in monsoon regions. Wang (2004) has found LS organization of MCSs to be commonplace over the northern South China Sea during the 1998 South China Sea Monsoon Experiment (SCSMEX).

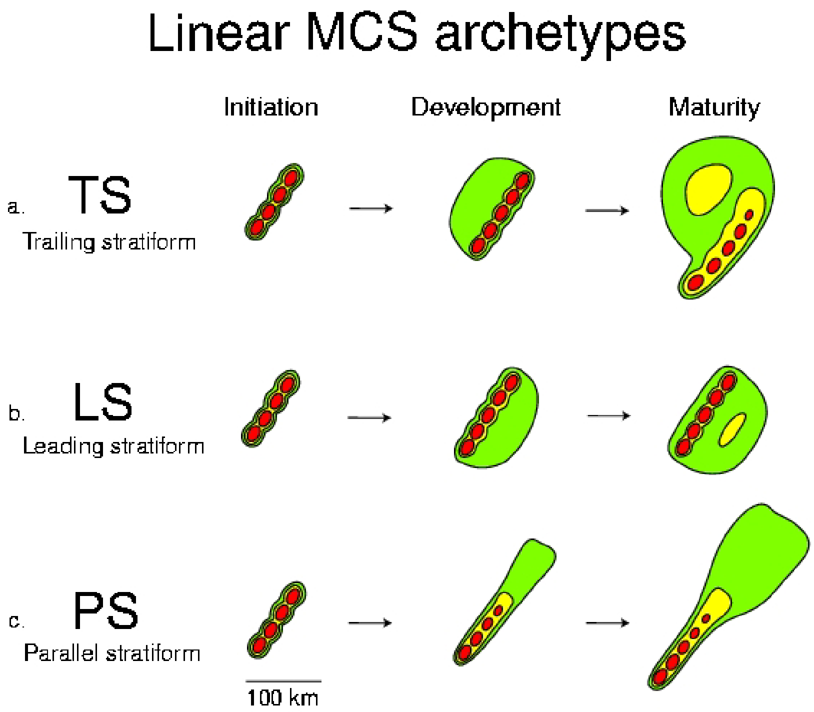

Figure 2. Three modes of organization of MCSs in the central United States (Parker and Johnson 2000).

TS systems normally propagate rapidly (~10-15 s^{-1}) and, as such, produce brief, heavy rainfall but rarely flash floods. LS and PS systems, on the other hand, can move more slowly and are often implicated in flash floods as a result of slow-moving, “training”, or back-building cells. An example of an LS system is shown in Fig. 3 (Pettet and Johnson 2003). The vertical structure is generally the mirror image of a TS system (Zipser 1977; Houze *et al.* 1989), with stratiform precipitation in advance of the convective line.

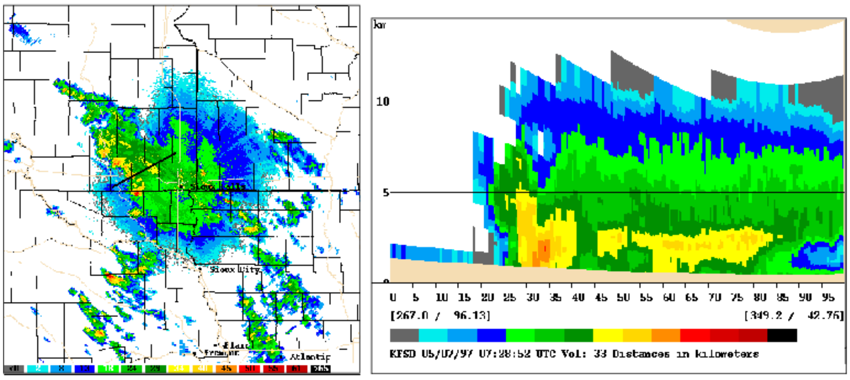

Figure 3. A leading-stratiform (LS) MCS over southeast South Dakota on 7 May 1997. System was moving eastward (left to right) and stratiform precipitation with bright-band in radar reflectivity was ahead of convective line (Pettet and Johnson 2003).

The heavy precipitation areas within the Meiyu (China), Baiu (Japan), and Changma (Korea) frontal zones during the summer monsoon are comprised of MCSs (e.g., Ninomiya *et al.* 1988a,b). Case studies of such systems have been reported in the literature, but there have yet to be climatological studies of the radar-inferred precipitation structures.

The convective and stratiform regions of MCS exhibit contrasting heating profiles, as shown in Fig. 4. The sharp increase with height of the heating and midlevels in the stratiform region produces a positive potential vorticity (PV) anomaly in the midtroposphere, leading to the generation of a midlevel mesoscale convective vortex (MCV). MCVs appear to be common over China in the Yangtze Valley during the Meiyu season, potentially contributing to long-lived precipitation systems, heavy rainfall, and flash floods (e.g., Akiyama 1984a,b).

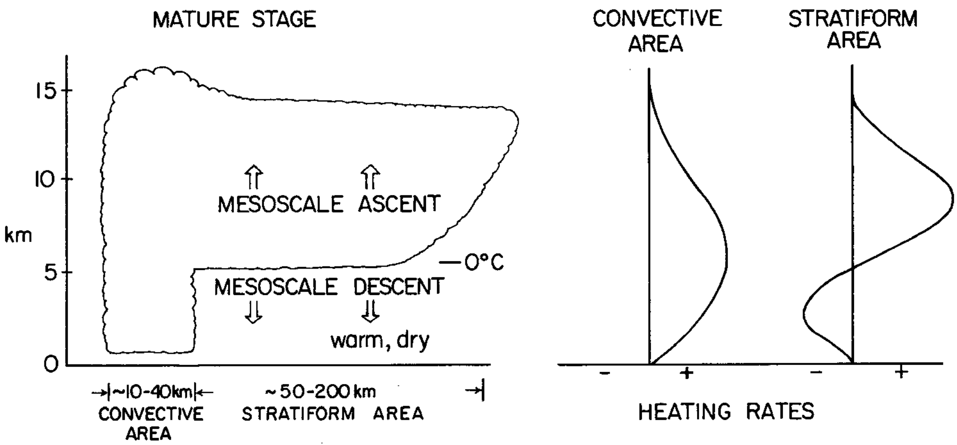

Figure 4. Idealized mature stage of MCS illustrating convective and stratiform precipitation areas along with associated heating profiles (Johnson 1986).

A climatology of the relative contributions of convective and stratiform precipitation regions to total rainfall has been carried out by Schumacher and Houze (2003) using the TRMM precipitation radar (PR). Their results (Fig. 5) show a wide variation in the stratiform rain fraction over the globe, with greatest amounts (50-60%) over the central and western Pacific and Indian Ocean. Low stratiform rain fractions (20-30%) are observed over Africa, parts of the maritime continent, and the Caribbean. This variability is not fully understood, but is likely related to the differing instability and wind shear in the different locations.

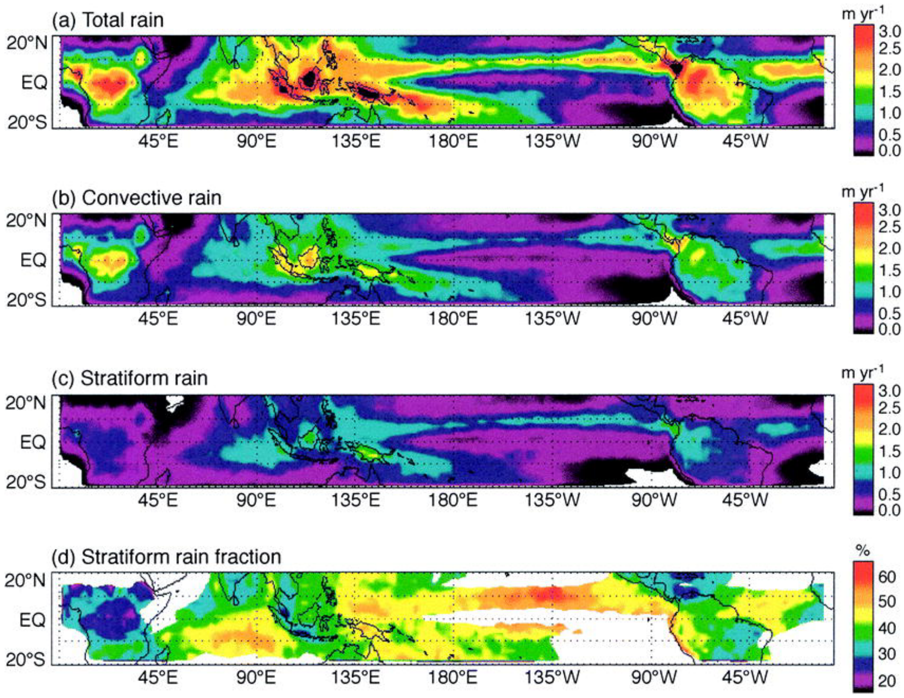

Figure 5. TRMM PR (a) total rain, (b) convective rain, (c) stratiform rain, and (d) stratiform rain fraction based on 2.5° grid averages for 1998-2000 (Schumacher and Houze 2003).

2. Topographically Forced Local Circulations

Throughout the world's monsoon regions, topography has a significant impact on local weather and precipitation (Riehl 1954; Ramage 1971; Ding 1994). In Fig. 1, the heavy rainfall at the foot of the Himalayas and adjacent to mountainous coastal regions is linked to topographic effects. Recently, Chang *et al.* (2004) have studied the relationship of rainfall to monsoon flow and topography over Indochina and the maritime continent using TRMM PR and QuikSCAT data. A map of topography over this region along with DJF and JJA QuikSCAT winds is shown in Fig. 6. Over most of the region there is a marked seasonal reversal of the flow. During boreal winter there is onshore flow toward coastal mountain ranges in Viet Nam, Malaysia and along the east side of the Philippines. This onshore flow contributes to boreal winter monsoon rainfall maxima in these regions, as seen in Fig. 7. This figure shows DJF minus JJA TRMM PR rainfall and QuikSCAT winds. Positive (negative) anomalies indicate maximum precipitation in boreal winter (summer). The positive anomalies along the east coasts of Viet Nam, Malaysia, and the Philippines indicate the maximum rainfall there in boreal winter. On the other hand, the negative anomalies off the west coasts of Myanmar, Cambodia, and the Philippines indicate maximum rainfall there during boreal summer in association with southwesterly monsoon flow (the reverse of that shown in Fig. 7). These results emphasize the important role of topography on precipitation distributions in the Asian monsoon region.

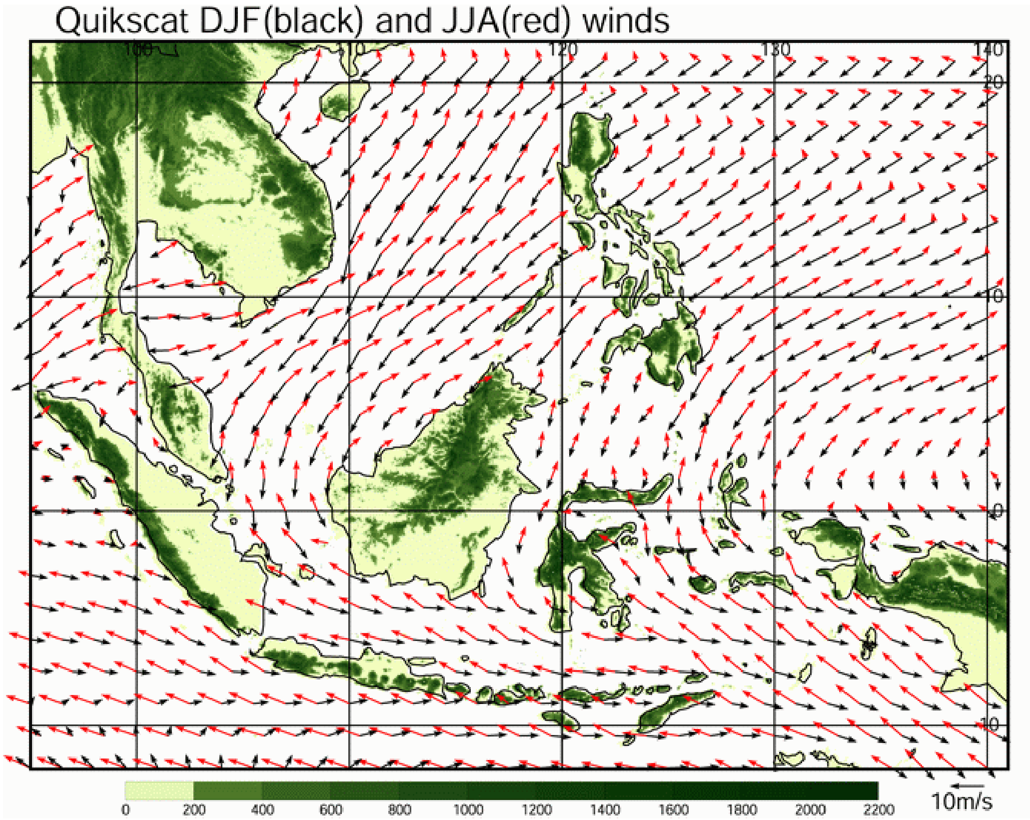

Figure 6. Mean QuikSCAT wind for DJF (black) and JJA (red), and topography (m), from Chang *et al.* (2004).

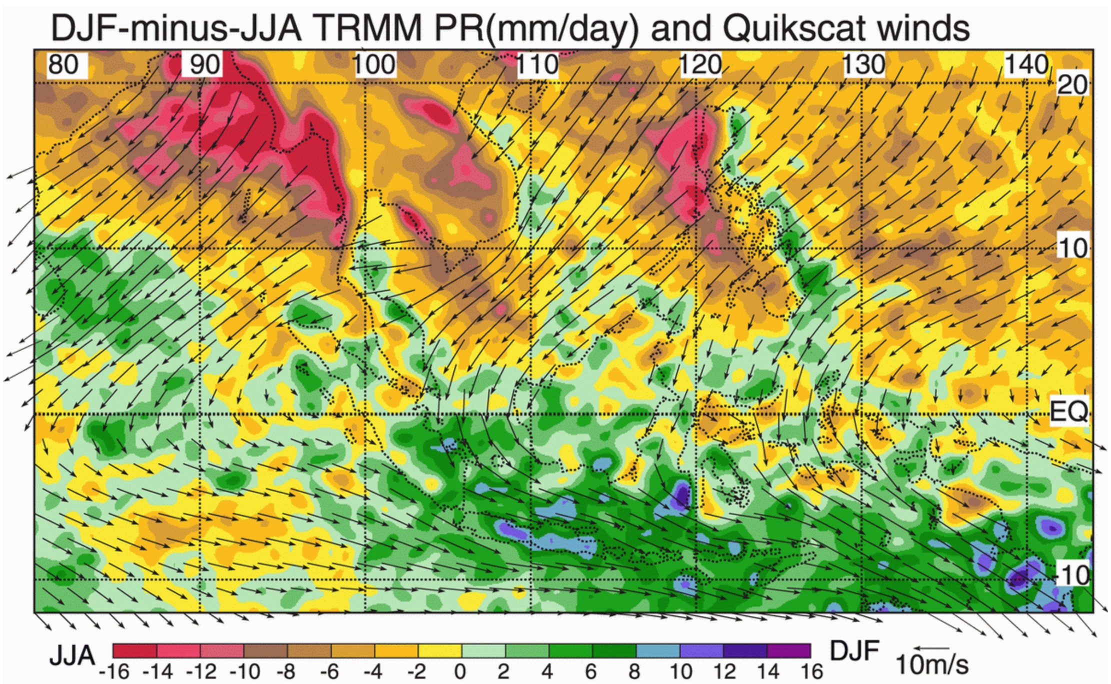

Figure 7. Differences of TRMM PR rainfall and QuikSCAT winds between boreal winter and boreal summer (DJF minus JJA). Warm (cool) colors indicate more rainfall in boreal summer (winter). From Chang *et al.* (2004).

Although it is not obvious from Figs. 1 or 7, much of the heavy coastal rainfall occurs just offshore rather than over the windward slopes of the coastal ranges. This behavior has been noted and studied for the heavy rain along the coast of western India upstream of the Western Ghats by Grossman and Durran (1984), Smith (1985), and Ogura and Yoshizaki (1988). The modeling study of Ogura and Yoshizaki suggests that the positioning of the heaviest rainfall just offshore is dependent on the strong vertical wind shear (low-level westerlies and upper-level easterlies) and strong surface fluxes over the ocean. Similar flow reversals occur during the boreal summer monsoon off Myanmar and the west coast of the Philippines, possibly helping to explain the similar behavior in those regions. Other possible factors for the offshore precipitation are upstream blocking of the low-level flow, land-breeze effects, and coastal frictional convergence.

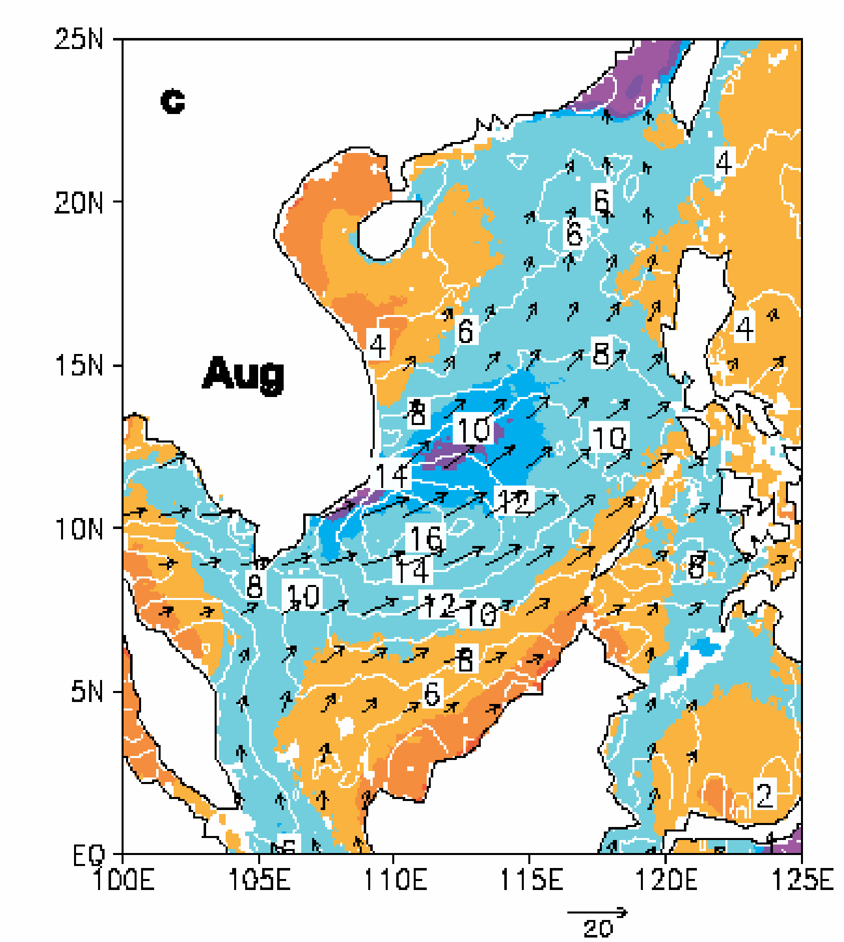

Figure 8. AVHRR 1985-1999 SST climatology (color shade in °C), along with QuikSCAT wind vectors and stress magnitude (contours in 10-2 N m^{-2}). From Xie *et al.* (2003).

There are a multitude of other mesoscale topographic effects in the monsoon regions of the world affecting local precipitation patterns. For example, flow blocking by Taiwan during the summer monsoon often leads to a low-level jet through the Taiwan Straits northwest of the island and lee vortices downstream, both of which can affect precipitation patterns (e.g., Chen and Yu 1988; Wang and Chen 2002). A particularly complex topographic effect reported by Xie *et al.* (2003) concerns the impact of the Annam Cordillera (the north-south mountain range on the east coast of Indochina shown in Fig. 6) on the flow over the South China Sea. During the summer months the southwesterly monsoon flow impinging on the Annam Cordillera creates a strong low-level jet off the South Viet Nam coast (Fig. 8). This jet leads to coastal upwelling of cool water, which is enhanced by Ekman upwelling due to the cyclonic curl of the wind sress on the north side of the jet. Figure 8 shows the coolest water displaced just north of the jet axis. The development of this cold filament in midsummer disrupts the summer warming of the South China Sea and causes a pronounced semiannual cycle in the SST. There is considerable interannual variability in this cold filament, for example, it did not develop during the 1998 SCSMEX year.

3. Diurnal Cycle

The diurnal cycle of the flow and precipitation is a dominant feature of the monsoons. On the large scale the Tibetan Plateau generates significant diurnally varying circulations, vertical motion, and diabatic heating features (Luo and Yanai 1983; Nitta 1983; Krishnamurti and Kishtawal 2000). On the mesoscale there are local land and sea breezes, and mountain/valley circulations that influence precipitation patterns over the monsoon regions of the world. During the 1978 Winter MONEX, the diurnal cycle of convection off the north coast of Borneo was studied using radar and sounding data. Houze *et al.* (1981) documented the development of nocturnal MCSs off Borneo, arguing they were a result of low-level convergence of the nighttime land breeze with the northeast monsoon flow (Fig. 9). The MCS begins as a group of convective cells near the coastline and later expands to a several hundred km scale system with both convective and stratiform components, later dissipating after sunrise as the sea breeze develops.

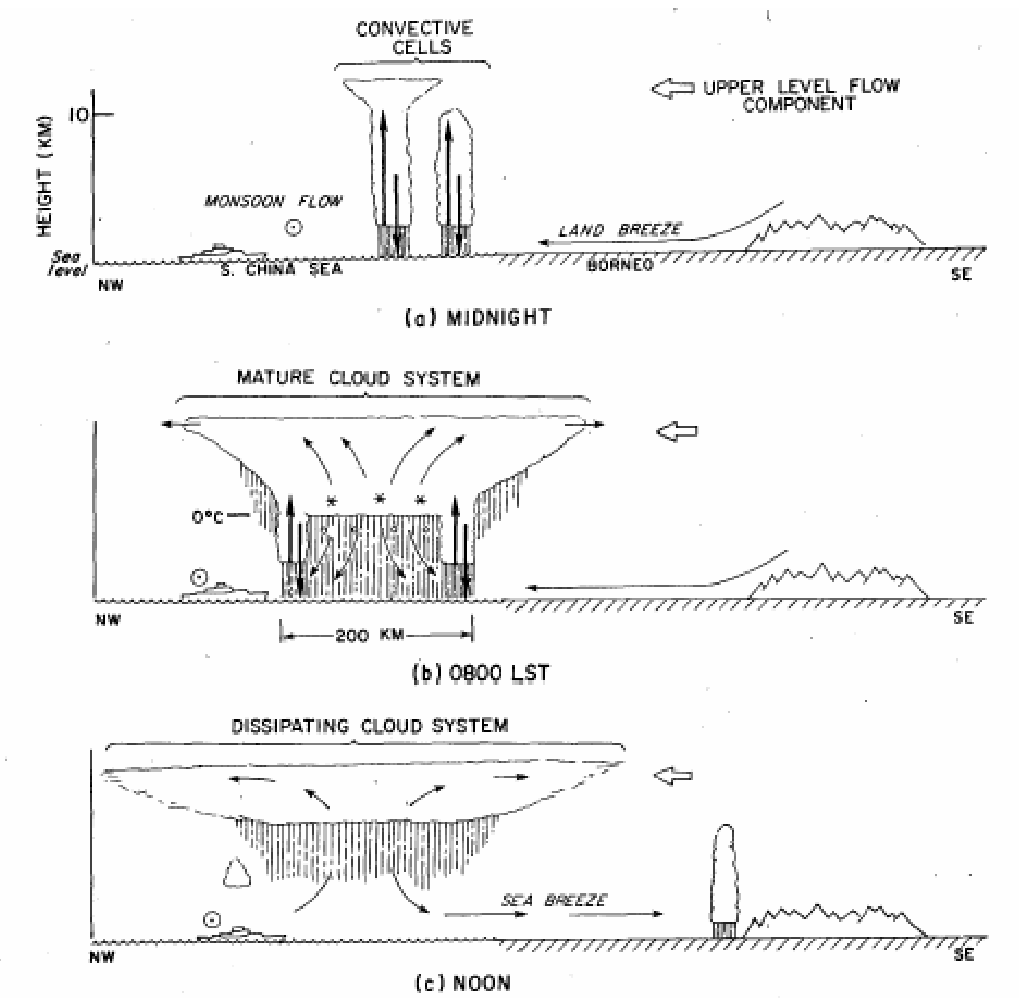

Figure 9. Schematic of the development of diurnally generated precipitation feature off the coast of Borneo (Houze *et al.* 1981).

In a study of convection over Taiwan during the Taiwan Mesoscale Experiment (TAMEX), Johnson and Bresch (1991) suggested that the land breeze flow at night was augmented by evaporation of the previous evening's precipitation over the interior elevated terrain. Mapes *et al.* (2003) proposed that the land breeze by itself was inadequate to account for nocturnal convection offshore as illustrated in Fig. 9. He argued that thermally forced gravity waves (produced by elevated terrain and propagating at about 15 m s^{-1}) are an essential part of the process, and that they produce a warm anomaly offshore during the daytime, thereby capping convection, while a cooling is produced at night, thus allowing convection to develop. This process is illustrated in Fig. 10.

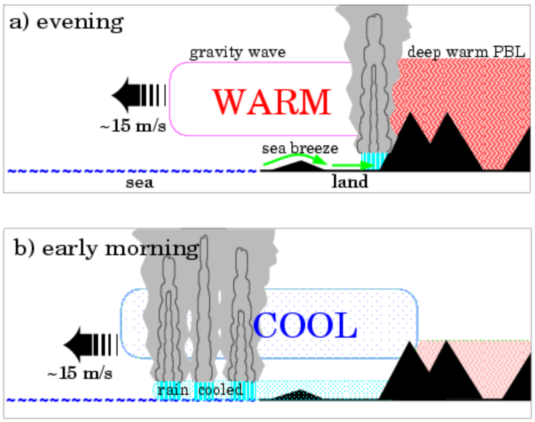

Figure 10. Illustration of the propagation of a thermally forced gravity wave by heating over elevated terrain and the initiation of offshore nocturnal convection (Mapes *et al.* 2003).

Understanding the diurnal cycle of convection in coastal environments is important because so much precipitation occurs in there and global models do not properly represent the diurnal cycle of convection (e.g., Yang and Slingo 2001). In the region of the Asian monsoon, satellite data indicate southward propagation of precipitation systems from India over the Bay of Bengal. This propagation is evident in a time-latitude diagram (Fig. 11) of brightness temperatures over the Bay of Bengal (Webster *et al.* 2002). Precipitation systems (inferred from the cold cloud tops) can be seen propagating all the way from the India coast near 20° N to the equator at times. Radar data from the R/V *Ron Brown* in the Bay of Bengal indicate that the convection associated with the diurnal signal has characteristics of squall line systems described in Section 1.

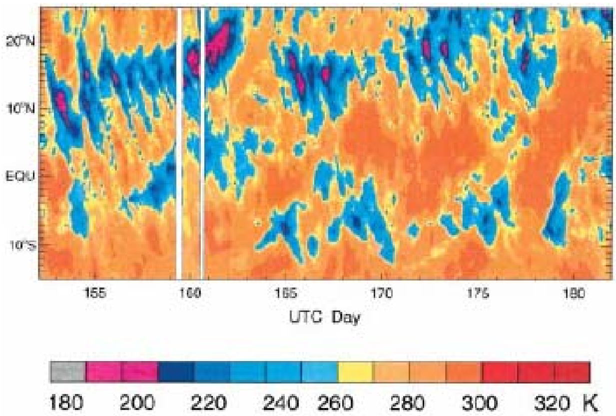

Figure 11. Time-latitude diagram of brightness temperatures (scale at bottom) from the European Space Agency METEOSAT-5 from 85 to 90° E for June 1999. From Webster *et al.* (2002).

4. Jets

As mentioned earlier, the topography of the monsoon regions often contributes to flow deflection or blocking and low-level jets. The low-level jet through the Taiwan Straits is one example. Low-level jets can also develop in response to boundary layer nocturnal cooling and an associated inertial oscillation, as observed in the African monsoon region (Blackadar 1957) and elsewhere. To illustrate the low-level jet (LLJ), consider the findings from SCSMEX shown below. The mean meridional wind component over the northern South China Sea at 22°N (Fig. 12) indicates peaks near 850 hPa at 02 and 08 L.

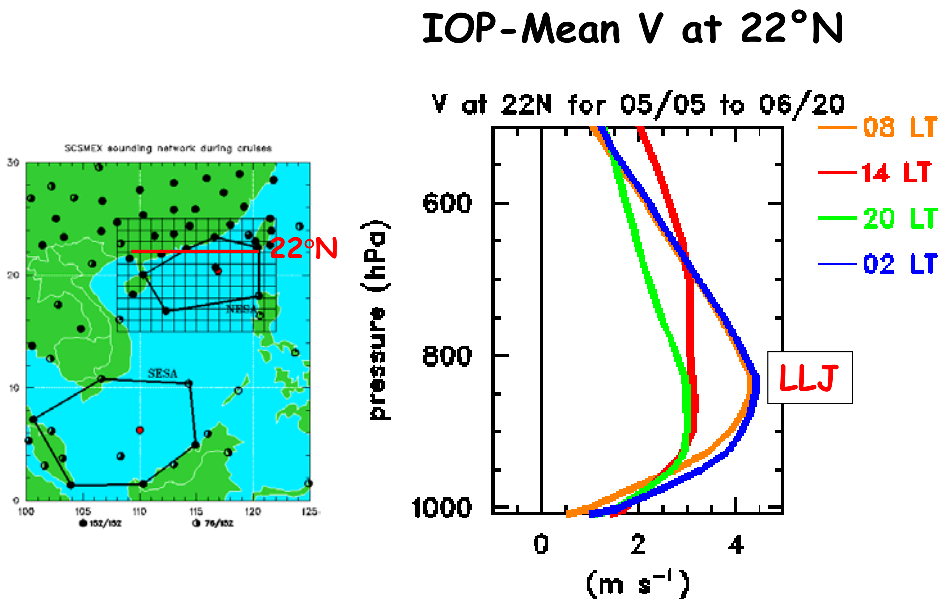

Figure 12. (Left panel) SCSMEX sounding arrays. Meridional wind is computed along line at 22°N. (Right panel) Meridional wind profiles at 02, 08, 14, and 20 L for period 5 May – 6 June.

The nocturnal wind maxima are indicative of an inertial oscillation. That such an oscillation is indeed occurring is evident from Fig. 13. A clockwise turning of the wind can be seen with maximum amplitudes at both Hong Kong and Dongsha Island at 08 L. The amplitude of the ageostrophic wind oscillation (~ 1 m s^{-1}), is considerably less than the ~5 m s^{-1} found over the summertime central United States (Whiteman *et al.* 1997), but it is not insignificant. Over the United States and in other regions the nocturnal LLJ has been linked with a nocturnal precipitation maximum, and the associated propagation of convective systems (e.g., Carbone *et al.* 2002). The linkage between the LLJ and precipitation along the south coast of China is not so clear, as illustrated in Fig. 14. In the afternoon, the sea breeze produces an afternoon maximum of convection over land. The 850-hPa LLJ is weakest at this time. In the early morning, observations show a precipitation maximum along the coast just offshore, which may be related to the land breeze, the gravity-wave mechanism of Mapes *et al.* (2003), or a combination of both. The 850-hPa LLJ is strongest at this time, but its linkage to the development of coastal convection is not clear. Once developed, the coastal convection propagates southward away from the shore, as shown in Fig. 15, similar to propagation over the Bay of Bengal seen in Fig. 11.

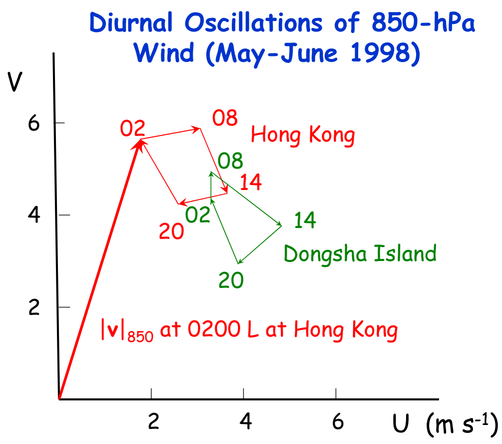

Figure 13. Diurnal wind oscillations at Hong Kong and Dongsha Island during SCSMEX.

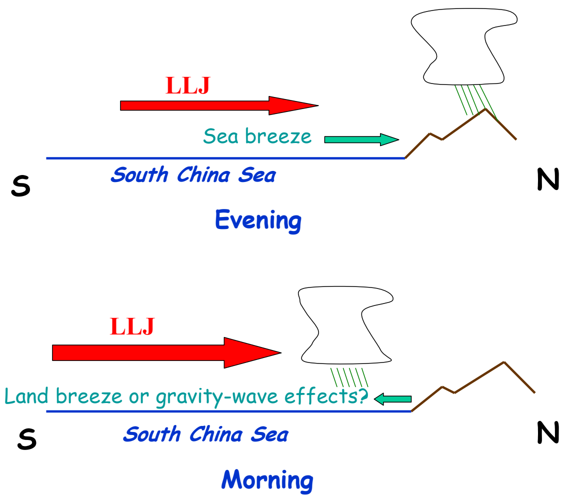

Figure 14. Schematic depiction of the diurnal cycle of convection over the northern South China Sea.

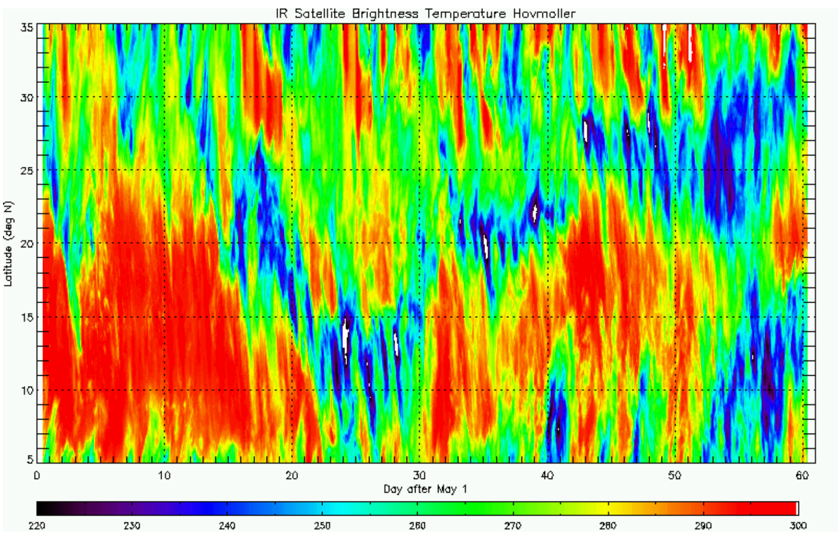

Figure 15. Time-latitude plot of IR brightness temperatures averaged over the South China Sea between 110 and 120°E for 1 May to 30 June 1998.

5. Surface-Atmosphere Interactions

Throughout the monsoon regions of the world surface exchanges represent important components of both the forcing of and response to the monsoon system (Ding 1994). For example, studies have shown that strong surface sensible heat flux over the Tibetan Plateau during the spring helps set the stage for the onset of the Asian summer monsoon by heating the upper troposphere, thereby contributing to an eventual reversal in the north-south temperature gradient (e.g., Flohn 1968; Luo and Yanai 1984; Li and Yanai 1996). After the summer rains begin, diabatic heating contributes further to this reversal and the overall energetics of the monsoon circulation.

Vigorous air-sea exchanges over the Arabian Sea and Indian Ocean have significant effects on the rainfall distribution over India as well as on the upper ocean. A prominent feature of the Indian summer monsoon is the abrupt cooling of the Arabian Sea following the onset of strong southwesterly flow in June. This phenomenon can be seen in Fig. 16, a time series of surface mooring data from the west-central portion of the Arabian Sea (Rudnick *et al.* 1997).

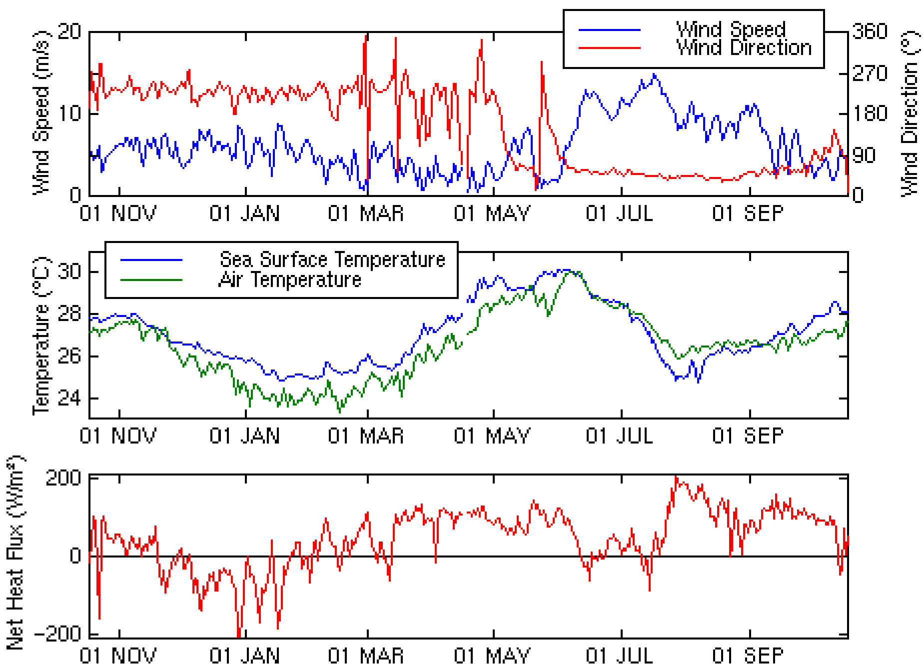

Figure 16. Daily averages of wind speed and direction (direction toward which wind is blowing), sea surface and air temperatures, and net downward heat flux as measured on the northern SIO mooring (Rudnick *et al.* 1997).

The sudden onset of strong southwesterlies around June 1 is accompanied by a sharp drop in the SST and air temperature. The SST-air temperature difference decreases to near zero after onset and there is a period of negative downward heat fluxes, primarily due to latent heat losses from the strong winds. The strong low-level jet over the Arabian Sea leads to a pattern of coastal upwelling (north of the jet axis) and downwelling (south of the jet axis) wind an overall southward Ekman transport (Fig. 17). This upwelling contributes to the Arabian Sea cooling, and it also brings nutrient-rich water to the surface, supporting increased productivity in the upper ocean.

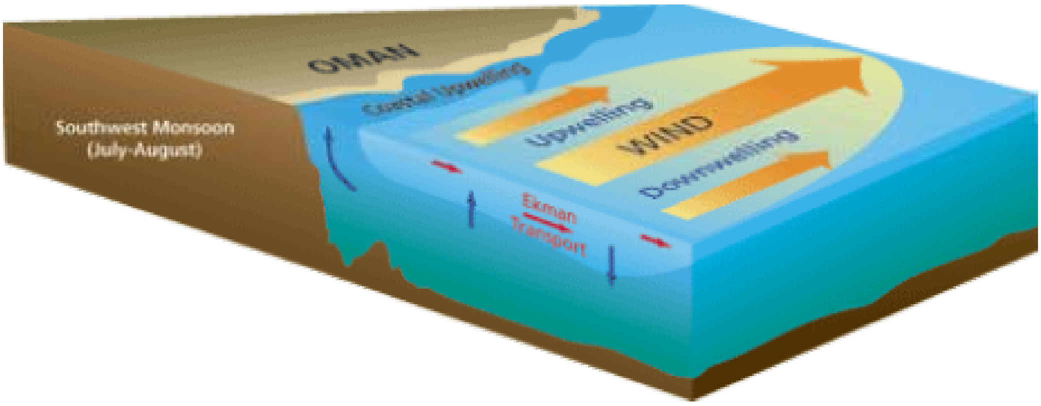

Figure 17. Schematic cross section of the upper ocean dynamical response to the strong southwesterlies of the Indian summer monsoon (Honjo and Weller 1997).

During the Asian winter monsoon, cold air often streams off the east coast of Asia, leading to strong sensible heat fluxes over the bordering oceans. The boundary layer over the East China Sea during cold-air outbreaks was sampled during the 1975 Air Mass Transformation Experiment (AMTEX). A visible satellite image of the cloud fields associated with a cold air outbreak over the East China Sea is shown in Fig. 18. Narrow cloud lines are seen to expand in scale to closed cellular patterns downstream. Surface sensible heat fluxes in these cold air outbreaks can reach 1200 W m^{-2} (Agee 1984), thus having a dramatic impact on the downstream circulation.

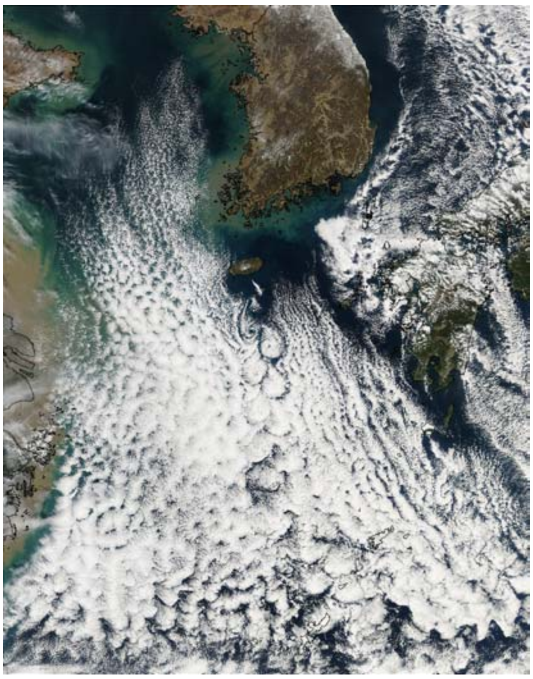

Figure 18. Aqua satellite image of cloud lines, closed cells, and vortex streets over the East China Sea on 15 January 2003 during a cold air outbreak.

Cold surges off the Asian continent were also studied during the 1978 Winter Monsoon Experiment (WMONEX). In addition to the large-scale impacts of cold surges (Lau and Chang 1987), there are regional impacts over the South China Sea and its surroundings. A plot of the temperature and streamlines at 900 hPa during the cold surge of 11 December 1978 is shown in Fig. 19 (Johnson and Zimmerman 1986). The data, obtained from aircraft dropsonde measurements, show a strong gradient in the 900-hPa temperature and strong cold advection over the northern South China Sea. In the 24-h period ending at 1200 UTC on 11 December, northeasterly winds accelerated over the entire span of the South China Sea and a cooling of up to 8°C occurred near the south coast of China.

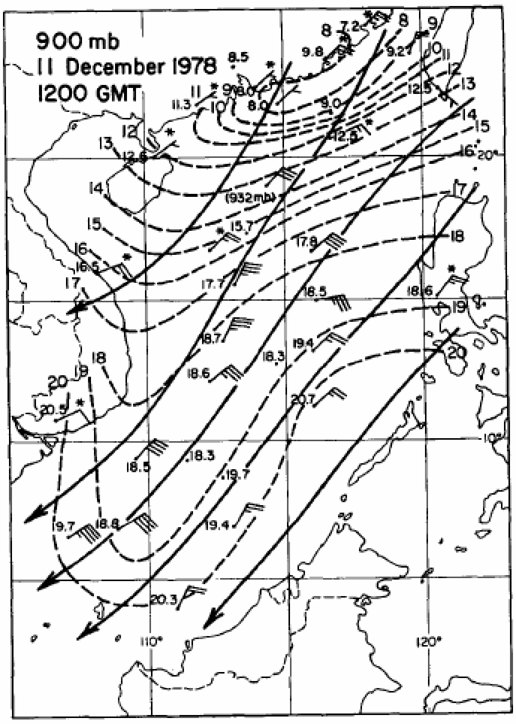

Figure 19. Streamlines and isotherms (°C) at 900 hPa at 1200 UTC on 11 December. Asterisk indicates winds at 850 hPa. Observations over water (other than from island stations) are based on aircraft dropwindsonde (WP-3D and Electra) and flight level (Hong Kong Islander) data (Johnson and Zimmerman 1986).

6. Coastally Trapped Disturbances

The (northward) southward movement of cold air to the west (east) of mountain barriers in the Northern Hemisphere contributes to sudden changes in local weather conditions in monsoon regions. These disturbances often exhibit characteristics of coastally trapped gravity waves, Kelvin waves, or Rossby waves (Skamarock *et al.* 1999), and their dynamics is still a subject of investigation. Douglas and Leal (2003) have used sounding data from the west coast of Mexico to demonstrate that Gulf of California surges during the Southwest Monsoon have characteristics of gravity currents (coincident sharp wind shift, temperature drop, and pressure rise in the lowest km). Cold fronts moving southward past Taiwan also exhibit properties of coastally trapped disturbances, with rapid propagation on the east side of the mountainous island, as illustrated in Fig. 20 (Chen *et al.* 2002). This rapid southward propagation along the east side of the barrier can be explained in terms of the shallow water equations. For a north-south barrier, the meridional component of the wind is given by $-fv = -g\,\partial h/\partial x$, where h is the free surface height. As the easterly geostrophic flow piles up cool air along the east coast of Taiwan, $\partial h/\partial x$ becomes negative, yielding $v < 0$ or a northerly flow. Similar effects are seen along the east coast of China as cold front surge southward just along the coastline. Cold fronts passing

Taiwan also display gravity current characteristics on the west side of the island, which impacts the properties of frontal convection there (Trier *et al.* 1990). Farther to the south as the cold surges move equatorward, their gravity-current characteristics diminish and their propagation is explained more by gravity-wave dynamics (Chang *et al.* 1983).

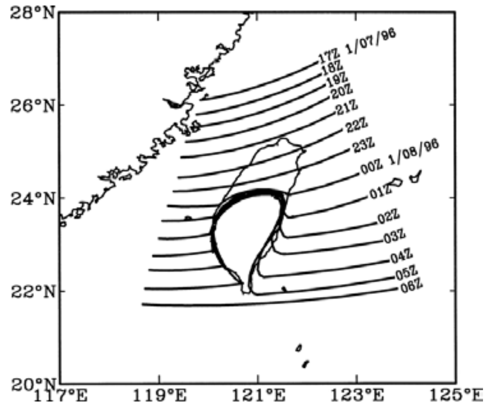

Figure 20. Locations of cold front at hourly intervals on 7-8 January 1996 (Chen *et al.* 2002).

7. Mesoscale Instabilities

Numerous mesoscale instabilities exist throughout the monsoon regions, many of which influence cloud and precipitation patterns. In the boundary layer, shear-modified Rayleigh instability, inflection-point instability, and Ekman layer instability may help explain the cloud streets and closed cell patterns seen in Fig. 18. Kelvin-Helmholtz (vertical shear) instability is prevalent in many regions and accounts for closely spaced (~10 km) banded structures in precipitation systems in sheared environments. Larger-scale banded structures in Meiyu or Baiu precipitation systems which have some baroclinicity may be associated with conditional symmetric instability (Bennetts and Hoskins 1979) or slantwise convection (Emanuel 1983). Inertial instability has been attributed by Toyoda *et al.* (1999) to explaining a series of anticyclonic vortices in a cloud band over Japan, where the instability is envisaged as arising from the vertical transport of momentum within deep convection. Cho and Chen (1995) have presented a theory for Meiyu frontogenesis that invokes Conditional Instability of the Second Kind (CISK), arguing that in contrast to the traditional Ekman layer CISK theory, boundary layer convergence is partly induced by friction and partly by latent heat release in deep convection.

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II. Synoptic Processes in Monsoons *(by Ding Yihui)*

1. Processes Relating to Monsoon Onset

The onset of the Asian summer monsoon is a key indicator characterzing the abrupt transition from the dry season to the rainy season and subsequent seasonal march. Numerous investigators have studied this problem from the regional perspectives. It is to some extent difficult to obtain a unified and consistent picture of the climatological onset dates of the Asian summer monsoon in different regions due to differences in data, monsoon indices and definitions of monsoon onset used in these investigations. Recently, Wang and Lin (2002) have identified two phases in the evolution process of the Asian summer monsoon. The first phase or the onset phase begins with the rainfall surges over the South China Sea in mid-May, which establishes a planetary-scale monsoon rainband extending from the South Asian marginal seas (the Arabian Sea, the Bay of Bengal, and the SCS) to the subtropical western North Pacific (WNP). The second phase of the Asian monsoon onset is characterized by the synchronized initiation of the Indian rainy season and the Meyu/Baiu in early to mid-June. The peak rainy seasons tend to occur primarily in three stepwise phases, in late June over the Meiyu/Baiu regions, the northern Bay of Bengal and the vicinity of the Philippines; in late July over India and northern China; and in mid-August over the tropical WNP (10°-22°N, 120°-160°E). The rainy season retreats southward in East Asia during late August and early September while the Indian summer monsoon and the WNPSM withdraws southward after mid-September. Ding (2004) has summarised the climatological dates of the onset of the Asian summer monsoon in different tropical regions based on various sources, with dividing the whole onset process into four stages: (1) Stage 1 (late in April or early in May): the earliest onset is often observed in the central Indochina Peninsula late in April and early in May, but in some cases, the onset may first begin in the southern part or the western part of the Indochina Peninsula. (2) Stage 2 (from mid to late May): this stage is characterized by the areal extending of the summer monsoon, advancing northward up to the Bay of Bengal and eastward down to the SCS. (3) Stage 3 (from the first dekad to second dekad of June): this stage is well known for the onset of the Indian summer monsoon and the arrival of the East Asian rainy season such as the Meiyu over the Yangtze River Basin and the Baiu season in Japan. (4) Stage 4 (the first or second dekad of July): the summer monsoon at this stage can advance up to North China, the Korean Peninsula (so-called Changma rainy season) and even North Japan.

Figure 21 presents an illustrative description of this onset process (Zhang *et al.*, 2004). During the first pentad of May (Figure 21(a)), the summer monsoon is established only over Sumatra. In the next two pentads (Figures 21(b)(c)), the tropical monsoon advances up to the land bridge, first establishing itself over the southwestern Indochina Peninsula and then expanding to the entire southern peninsula. During the pentad of May 16-20 (Figure 21(d)), the build-up of the summer monsoon is observed over the central Indochina Peninsula. At the same time, the onset location extends into the central and southern SCS, accompanied by a rainfall rate of >5 mm day^{-1} over the entire SCS. In the next pentad, onset expands quickly and almost covers the entire SCS (Figure 21(f)). On the other hand, the Asian summer monsoon also advances northwestward to the Indian monsoon region from the near-equatorial East Indian Ocean and the Indochina Peninsula starting from mid-May (Figure 21(d)). Earliest onset of the Asian summer monsoon in this region may be observed over the southern tip of the Indian subcontinent. In early June (Figures 21(g) (f)), the Asian summer monsoon rapidly advances northwestward, arriving in the central Indian subcontinent. Meanwhile, the onset over the Arabian Sea and the western coast of the Indian subcontinents is observed, due mainly to the enhancement of the cross-equatorial airflow off the Somali coast and the development of the onset vortex in the central and northern Arabian Sea (Krishnmaurti *et al.*, 1981; Ding, 1981). This date is generally believed to be normal onset dates for the Indian summer monsoon. So, the onset of the East-and Southeast Asian summer monsoon and the South Asian summer monsoon is closely interrelated in the context of the Asian summer monsoon. However, the earliest onset of the Asian summer monsoon occurs over the Indochina Peninsula and the SCS.

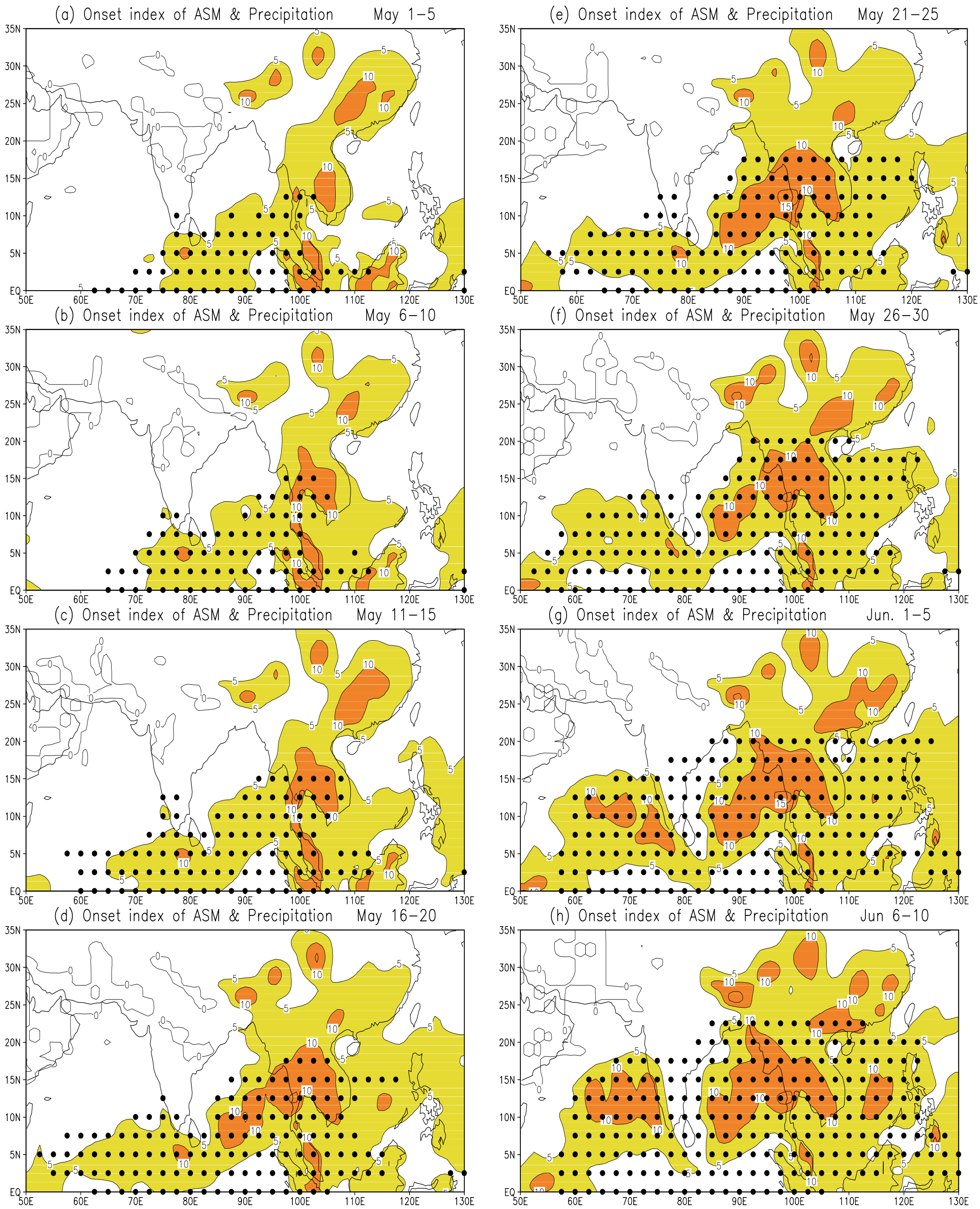


Figure 21. Climatological pentad-averaged precipitation rates (mm day-1) for the period from May1-5(a) to June 6-11(h) in sequence. Light and dark shadings indicate precipitation regions greater than 5mm day^{-1} and 10mm day^{-1}, respectively. The black dots represent the location of onset of the summer monsoon (Zhang *et al.*, 2004).

Similar to the onset vortex over the Arabian Sea, a twin cyclone is often observed to develop in the near-equatorial East Indian Ocean (near Sri Lanka) prior to the onset of the summer monsoon over Indochina Peninsula and the South China Sea. The equatorial westerlies are accclerated and extend northeastward into the Indo-China Peninsula and the northern SCS, thus greatly enhancing the low-level westerlies there. Therefore, a pre-condition of the first onset of the Indochina Peninsula and SCS summer~~s~~ monsoon is the development of a twin cyclone over equatorial East Indian Ocean. The climatological and case study both have shown that this pre-condition generally exists (Lau *et al*., 1998, 2000; Zhang *et al.,* 2001). For 13 cases of the first monsoon onset of the northern SCS out of 47 years from 1953 to 1999, the twin cyclones over the equatorial East Indian Ocean were always observed, although their developmental processes may have some differences. Figure 22 is the composite 850 hPa wind field for 13 cases of the first onset of the northern SCS summer monsoon, showing the existence of the twin cyclone (C1 and C2) straddling the equator over the tropical East Indian Ocean and significant acceleration of equatorial westerlies in this region, which can extend northeastward up to the northern SCS.

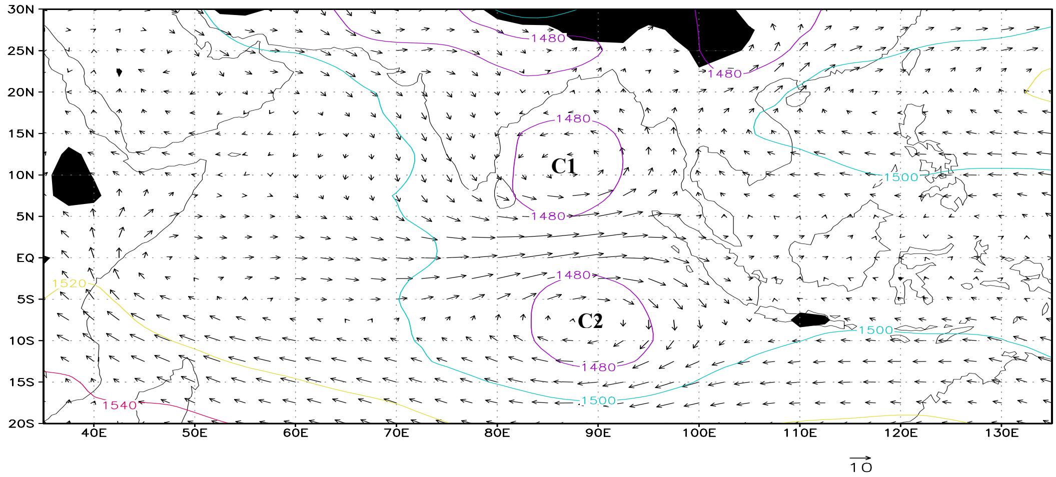


Figure 22. The composite 850 hPa streamline pattern averaged for 13 cases for one pentad before the onset of the summer monsoon over the northern SCS (a) and during the onset of the summer monsoon over the northern SCS (b). 13 cases used for average include the following years: 1957, 1968, 1969, 1972, 1973, 1981, 1982, 1983, 1984, 1992, 1993, 1997 and 1998 (Zhang *et al.*, 2001).

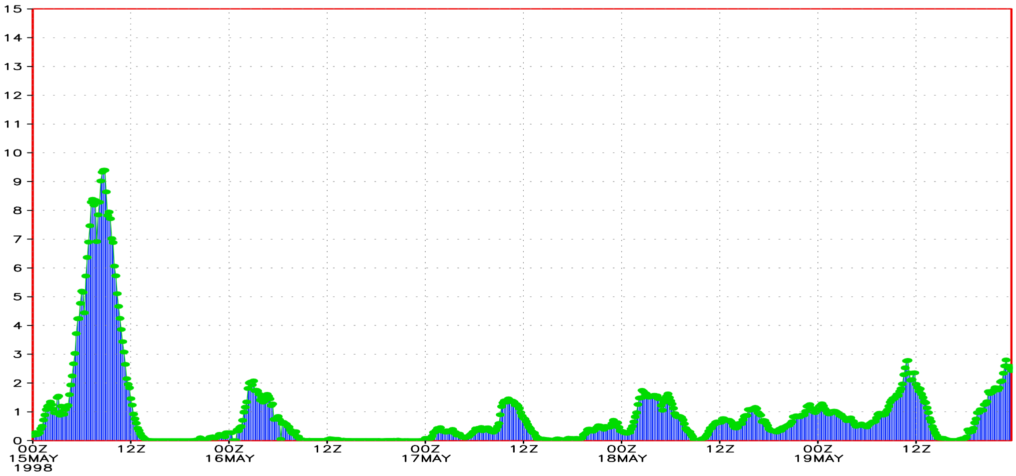
Another pre-condition for the first summer onset of Indochina Peninsula and the South China Sea (especially northern SCS) is the southward intrusion of strong cold air accompanied by Chang and Chen (1995) an obvious cold front. As pointed out by Chan *et al.* (2000), and Ding and Liu (2001) ,this process from mid-latitudes may be a triggering mechanism for the monsoon onset in the northern SCS, especially for rapid amplification of regional precipitation and convection in these regions. This condition is different from that of the Indian summer monsoon whose onset is basically forced by tropical impacts. Recently, Liu *et al.* (2002) suggest that the southward intrusion of midlatitude frontal systems is related to the downstream propagation of the Rossby wave excited by convective activity in Bay of Bengal.

One important indicator characterizing the onset of the Indochina Peninsula and the SCS is the precipitation and convection. Based on observations made by the dual-Doppler radars deployed in the northern SCS by SCSMEX, the organized convections in the monsoon trough were very vigorous after the onset, which often organized into meso-scale rainbands or meso-scale convective systems (MCSs). These MCSs are closely related to the synoptic processes and systems (Fgiure 23) (Ding, Li and Liu, 2004). In particular, the establishment of the low level monsoon trough and associated wind shear line during the monsoon onset provides the favorable synoptic and dynamic condition for formation and maintenance of MCSs. The latter will in turn create a significant feedback effect on large-scale circulation and synoptic system in which are embedded the MCSs, through vertical heat and moisture transport and release of latent heat. Johnson and Ciesielski (2001) have recently studied the characteristics of the onset of the summer monsoon over the northern SCS utilizing observation from the May – June 1998 SCSMEX. They have shown that the onset occurred in the northern SCS in mid-May with a rapid increase in deep convection over a week to ten-day period, thus producing the significant vertical transports of heat and moisture. During the undisturbed pre-onset period (May 6-12) (Figure 24), there is upper-level convergence, low-level divergence and deep subsidence, consistent with the mostly clear skies and high values of OLR. Q_1 (the apparent heat source) is negative at all levels with values in excess of -2.5K day^{-1} in the upper troposphere. The profiles of divergence and vertical motion during the onset (May 16-22) and post-onset periods (June 3-9) are dramatically different from the pre-onset period. Low-level convergence, upper-level divergence and strong upward motion occur during both periods. Deep convergence extends to 300-500 hPa, Q_1 and Q_2 (the apparent moisture sink) profiles are characteristic of deep convection during these periods, with a peak heating rate of 5 K day^{-1} located at about 400 hPa and a considerable separation of the two curves, suggesting more vigorous deep convection at that time. Thus, the intense heating can led to the surface pressure in the monsoon trough further to fall down and tropical low-level southwesterlies to monsoon trough was enhanced. This positive feedback process is also favorable for eastward retreat of the subtropical high out of the SCS region.

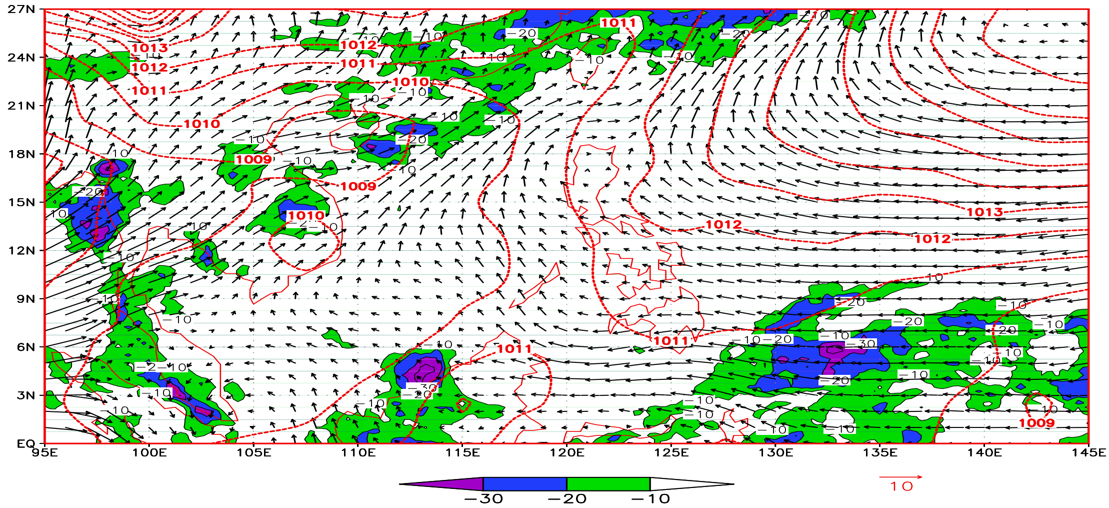
The onset of the monsoon is the most anxiously awaited weather singularity in the sub-continent as it heralds the rainy season and marks the end of the hot summer (Ding and Sikka, 2004). Over continental India, the drama of the onset of the South Asian summer begins first across Kerala coast, normally by 31 May (Ananthakrishnan and Soman,1988) when heavy rains lash the coastal state after the cross equatorial low level jet (LLJ) is established across the Somali coast into near-equatorial Arabian Sea. This phenomenon is usually accompanied by the formation of a mid-troposphere shear zone across central Bay of Bengal to SE Arabian Sea in which may be embedded a cyclonic vortex. The vortex may even intensify into a cyclonic storm either in the Bay of Bengal or SE Arabian Sea. The cyclonic storm forming in the SE Arabian Sea is known as the Monsoon Onset Vortex (Krishnamurti *et al.*, 1981) after the event which occurred on 11 June 1979 during the Summer MONEX year.

The vortex is formed to the north of the LLJ, in the zone of maximum cyclonic shear in the lower tropospheric zonal winds. In association with the northward movement of the vortex, the large scale monsoon current also advances northward along the West Coast of India. On the average, the onset vortex forms in nearly 50 percent of the years and in other years the onset is accompanied by either mid-tropospheric shear line or formation of an off-shore trough along 10 to 15^{0}N off the Kerala coast of India with an embedded weak low pressure area. Prior to the onset the cross-equatorial flow increases in strength, the moisture fields builds upto the mid-tropospheric level 7 to10 days in advance (Pearcl and Mohanty, 1984), the near-equatorial cloud band in the Arabian sea expands eastward and the upward motion field in the troposphere near the equator enhances which eventually moves northward with the advance of the monsoon.

Yin (1949) was the first to link the process of monsoon onset to the displacement of westerly troughs in the circumpolar westerlies and shift of STJ to the north of the Himalayan periphery. Pasch (1983) examined the monsoon onset over India during 1979 from the perspective of planetary scale circulation features and noted their gradual setting up prior to the formation of a transient disturbance off the Kerala coast. Murrakami and Ding (1982) have opined that the onset is related to the warming of the Eurasian region by diabatic heating. Yanai *et al.* (1992) have linked the onset to the effect of the Tibetan Plateau. Thus the onset of monsoon over India is linked to a combination of regional and planetary scale changes over the entire Indian Ocean region. There exists a variety in the linkages of the onset process with the seasonal developments or transitions in the regional and planetary scales features. Lack of uniqueness among them points to the initiation of chaotic dynamics with the formation of the weather system at Kerala coast which heralds the onset process.



(a)



(b)

Figure 23. (a) Retrieved precipitation per ten minute from dual-Doppler Radar during May15-19 in 1998, which was mainly resulted from meso-scale convective systems (MCS) (unit:mm.hr^{-1}); (b) 850 hPa wind field sea-level pressure (unit: hPa) and TBB (unit: 0C) during May15-19 in 1998. Dashed box : the region investigated in the northern SCS; bold black line: monsoon trough. (Ding, Li and Liu, 2004)

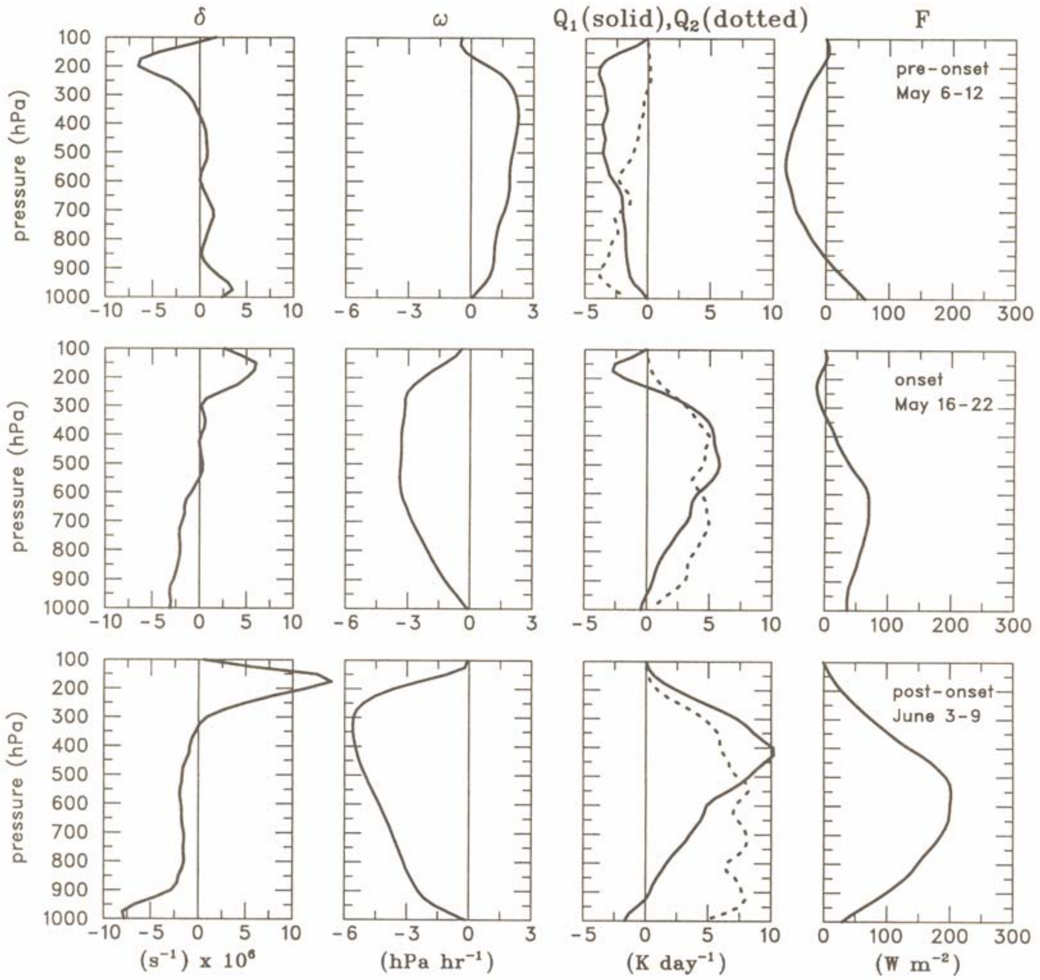


Figure 24. Vertical profiles of divergence (δ), vertical motion (ω), Q1, Q2 and total flux F for the undisturbed pre-onset period (top panel), the onset convective period (middle panel), and the post-onset convective period (bottom panel) (Johnson and Ciesielski, 2002).

2. Frontal Dynamics

The frontal dynamics have been widely studied in relation to different structures and the life cycle over the various regions over the world. A comprehensive review of these issues was provided by Keyser and Shapiro (1986). One core component of frontal dynamics is the interactions between the primary (geostropihic) and secondary (ageostrophic) circulations. In this review paper, the two-dimensional theory of forced secondary circulations in the cross-front plane developed by Sawyer and Eliassen is presented to elucidate the mechanism and feedback contributing to the evolution of upper-level fronts in relation to their setting within baroclinic waves. Theoretical and numerical examples of the formation of upper-level fronts in idealized two-dimensional flow are reviewed. In the three-dimensional case, the presence of along-front ageostrophic circulations superimposed upon the cross front ageostrophic circulations treated by the two-dimensional theory is discussed in terms of gradient wind. The relative contribution of along-front ageostrophic circulation to upper level frontogenesis is considered in the context of the results from three-dimensional β-plane channel models of baroclinse wave growth. The characteristics and dynamics of the coastal front along the southeast coast of the United States have been extensively examined. The coastal front is frequently a cyclogenesis site because it provides a narrow zone along the coast where the low-tropospheric convergence, warm-air advection, baroblinity and cyclonic vorticity are maximized (Bosart and Lin, 1984; Doyle and Warner, 1993). In East Asia, the coastal frontogenesis is sometimes observed (Chen and Dell'Osso, 1987).

The Meiyu/Baiu front is a unique frontal system in East Asia that occurs in the moist atmosphere during the East-Asian summer monsoon (June 12-July 10). Many works have been devoted to their characteristics, structures and dynamics (Ding, 1992; Chen, 2004; Ninomiya, 2004). The major Meiyu region is located in the Yangtze and Huaihe River basins, with the rainfall amount accounting for 30-50% of total summer rainfall. The Meiyu fronts occur in a high θse atmosphere. In mid-May to mid-June, the axis of maximum frequency of 850hPa Meiyu front in South China and Taiwan, indicating that the mean position of the Meiyu front is oriented approximately in an east-west direction extending from southern Japan to Southern China. The mean position shifts northward to Japan and central China in the Meiyu season of the Yangtze River valley (mid-June to mid-July), as a quasi-stationary front with an average lifetime of 8 days (Chen, 1988).

A conceptual model of the meso-scale cloud system family in the Meiyu-Baiu front is proposed in Figure 25 by Ninomiya (2004). In this model the cloud zone consists of a few cloud system families, each of which consists of two parts: a sub-synopticscale cloud system associated with a sub-synoptic-scale Meiyu-Baiu frontal depression (indicated by S), and a few meso-α-scale cloud systems (indicated by α). The latter are aligned along the trailing portion of the preceding sub-synoptic-scale cloud system. Cold lows and a midlatitude blocking ridge and the Pacific subtropical anticyclone all have strong influences on the Meiyu-Baiu cloud sytems. Rows of large and small arrows in Figure 25 indicate the 500-hPa and 850-hPa maximum wind zone becomes coupled with the shortwave trough in the Meiyu-Baiu frontal zone under the influence of the cold low over Siberia, leading to the development of a sub-synoptic-scale frontal depression. Subsequently, a few meso-α-scale cloud clusters form along the trailing portion of the preceding sub-synoptic scale cloud system. The Meiyu front affecting South China and Taiwan forms in the subtropical latitude, which is a distinct area from that for the formation of polar front in the Meiyu season. It resembles a semitropical disturbance with an equivalent barotropic warm core structure, a weak horizontal temperature gradient, a rather strong horizontal wind shear, and a positive low-level potential vorticity (PV) anomaly. Chen (2004) has summarized dynamic aspect of the Meiyu front for its early stage (South China and Taiwan). The CISK mechanism has been suggested for the Meiyu frontogenesis through the interaction between the PV anomaly and the convective latent heating. Some of the Meiyu fronts observed in TAMEX were characterized by the density-current type structure at the leading edge with a thermally direct circulation cell across the front. The frontal deformation occurs with different propagation speeds to the east and west of the Taiwan Central Mountain Range. The acceleration of highly ageostrophic flow of cold air in response to the building pressure gradient has been suggested to responsible for the faster moving front to the east of Taiwan.

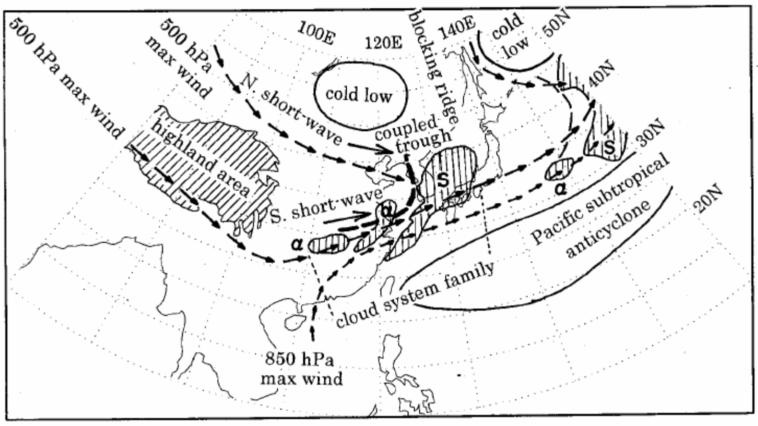


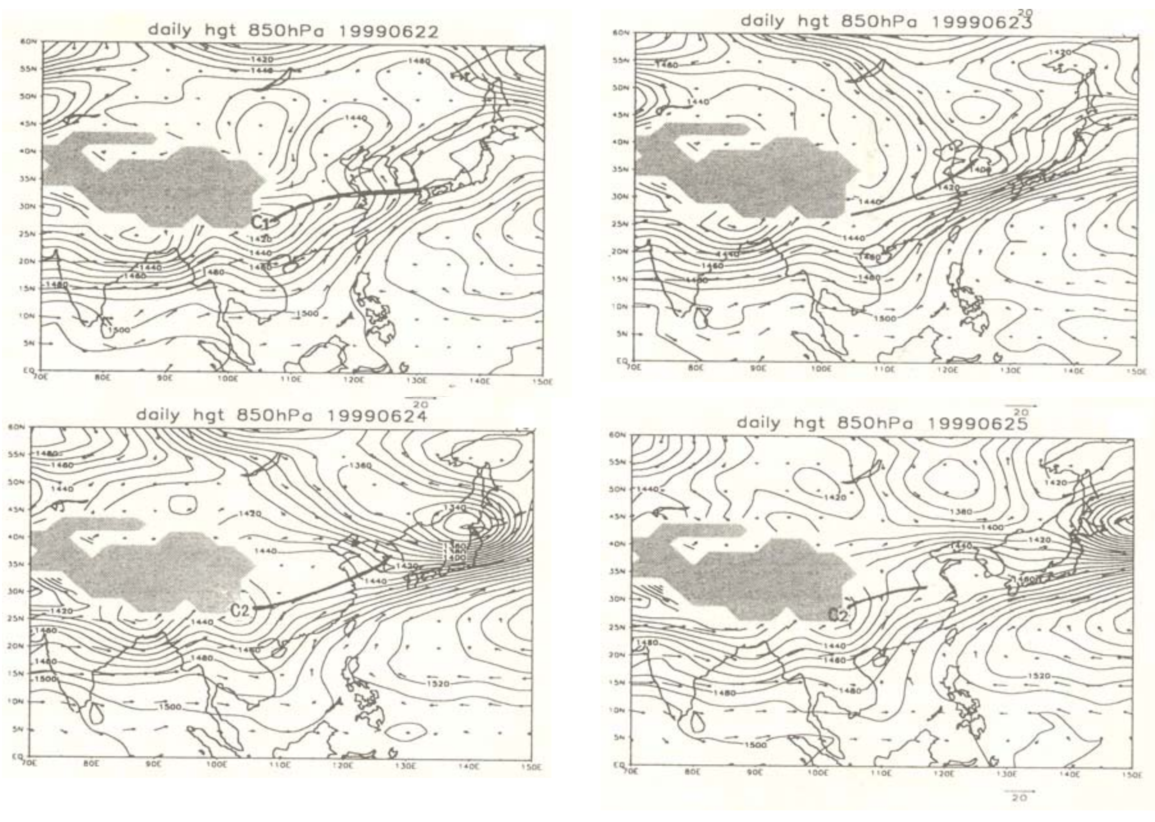
Figure 25. Conceptual model of the Meiyu-Baiu frontal cloud zone. (Ninomiya, 2004)

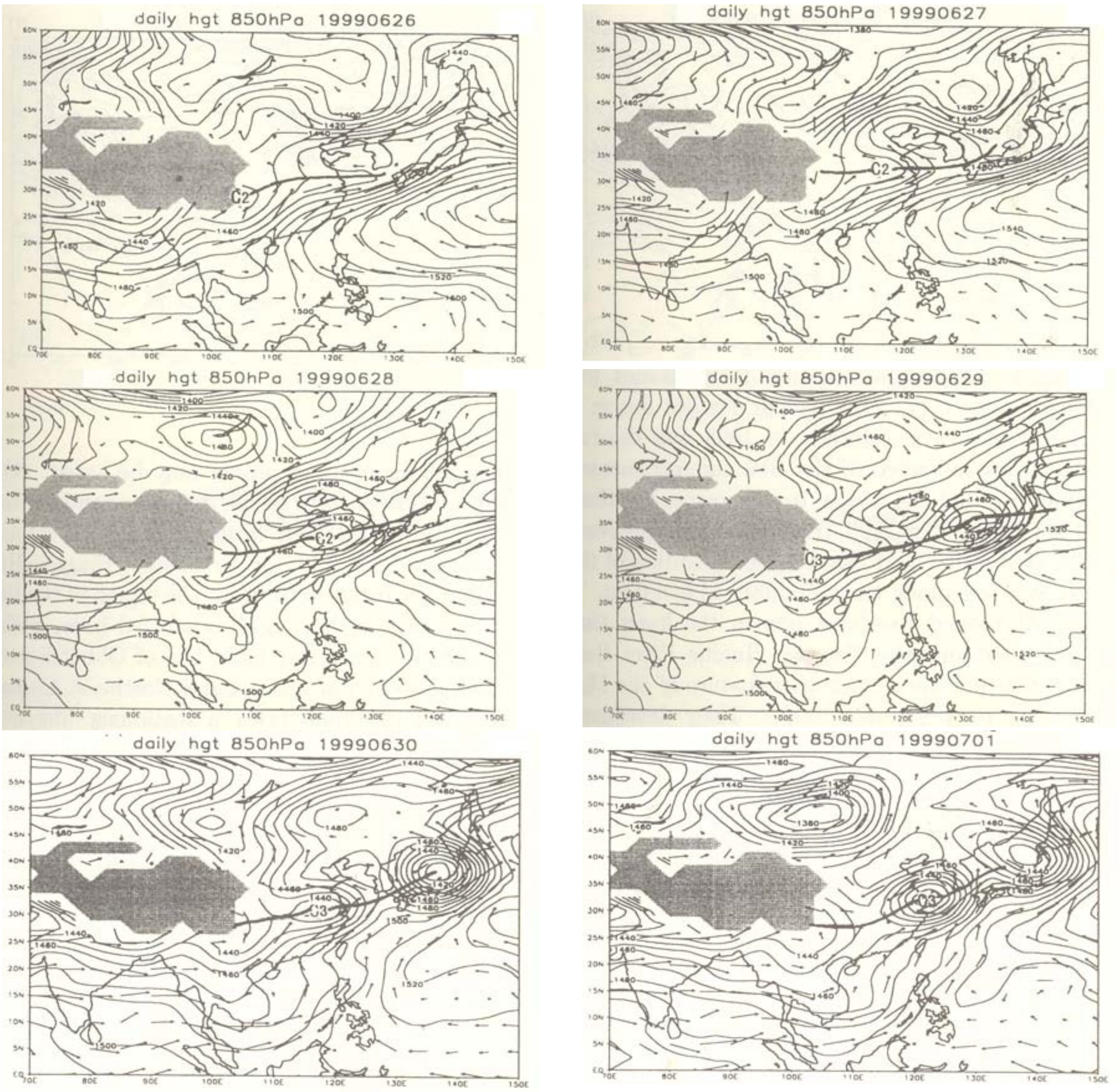
3. Topographically Influenced Circulations

The low-level vortices produced locally over the Tibetan Plateau and its sloping periphery are unique weather systems of a sub-synoptic scale in the East Asia monsoon region. These topographically generated or influenced circulation systems are among the most important rain-bearing systems in East Asia during spring and summer (Ding, 1992). Low-level vortices have their geographical origin in three areas: the southeastern part of the Plateau (called southwest (SW) vortices), the main body of the Plateau (called Plateau vortices) and the northern part of the Plateau (termed northwest (NW) vortices because of their location in the northwestern part of China) (Tao and Ding, 1981; Ding, 1994). These vortices can produce excessive rainfall after they leave the Plateau if they are supplied with ample moisture and are supported by a favorable large-scale environment. The genesis and development of these vortices, especially SW vortices, are closely related to activity of monsoonal airflows. Therefore, they may be considered to a certain extent as the result of interaction between the Plateau and monsoon. The Plateau vortices are obviously a weather system of a small scale, shallow depth, weak intensity, and short life cyclone. They are generated by a particular underlying surface. Once they move out of the Plateau, they generally dissipate rapidly due to changed conditions of the underlying surface. The Plateau-produced mesoscale vortices are often generated by organized mesoscale convection. During the summer, the Plateau receives shrong solar radiation, and the overlaying atmosphere often displays strong convective instability, even under the influence of a subsiding high pressure system. With strong solar radiation, every mountain peak acts as an isolated heat island which triggers convection if moisture is available (Reiter and Tang, 1984). If the convective activity persists long enough, a mesoscale cyclonic circulation can develp and, in turn, organize cumulus convection. In this case, the air over the Plateau is often convectively unstable ($\partial\theta e/\partial P>0$) at low level (Kuo, Cheng and Anthes, 1986). A maximum of θ_e exists over the Plateau. The center of the Plateau vortex corresponds very well to the maxima of θ_e, suggesting the warm core structure of the Plateau vortex. The unstable air stratification may be caused by strong sensible heating from the surface, especially over the central and western parts of the Plateau. Recent studies (Chen and Dell'Osso, 1984; Kuo *et al.*, 1986) suggest that both baroclinic instability and latent heat release associated with organized convection over the Plateau might play important roles in the generation of the Plateau vortices. A vortex often develops within an existing shearline that separates the warm, moist monsoon flow to the south from the midlatitude colder, drier air to the north. In fact, this Plateau-produced vortex is embedded within a sharp, short-wave trough. This vortex is also associated with strong convective activity throughout its life cycle. Chen and Dell'Osso (1984) also showed that the intensity of this vortex was significantly weakened when latent heat release was removed from the model.

In general, the SW vortex is defined as a 700 hPa closed cyclonic circulation over southwestern China, mainly over the western part of the Sichuan Basin. It is a low-level circulation system, often only visible on 850 and 700 hPa analyses. On the surface weather map, one may often observe a negative pressure tendency during 24 hours over the low-vortex region. In this sense, the SW vortex is also called the SW low vortex. Most of the time, the vortex forms and dissipates without intense development and the severe weather associated with it, because there is usually a lack of upper-level support and moisture. But if the synoptic conditions are favorable, the SW vortex may cause severe weather, especially heavy rainfalls. The SW vortex may provide strong orographic lifting to trigger convection and, consequently, a large amount of rainfalls on the steep topography surrounding the Sichuan Basin. Many cases may be exemplified, for example, the heavy rainfalls in the Sichuan Basin on 1-14 July of 1981 which have been extensively studied by numerous meteorologists (Chen and Dell'Osso, 1984; Kuo, Cheng and Anthes, 1986; Wang and Orlanski, 1987). This heavy rainfall took a large toll in human life and caused much damage to property. If the SW vortex moves eastward after its formation, it can cause heavy rainfall over the middle and lower Yangtze River Valleys. If the SW vortex take a northward track of movement, they can bring heavy rainfall to northern China. Figure 26 is a notable example of consecutive genesis, development and eastward movement of three SW vortices in the 1999 Meiyu season (Ding *et al.*, 2001).

From the synoptic viewpoint, the genesis and development of the SW vortex needs to meet two requirements: (1) the existence of a vigorous southerly airflow from the eastern slope of the Tibetan Plateau to the Sichuan Basin. It may play a dual role in the genesis of the SW vortex. Dynamically, this southerly wind produces "differential frictional effects", a mechanism first discussed by Newton (1956) in connection with Colorado cyclone formation, thus leading to the formation of a cyclonic circulation at low level. Thermally, the southerly wind may transport abundant warm, moist air into the eastern slope of the Plateau and the Sichuan Basin, providing the major moisture source for precipitation and the release of latent heat; (2) the necessary triggering mechanism. Most of the time, the low pressure troughs passing over the Tibetan Plateau may act as a triggering mechanism for the SW vortex. Chang *et al.*, (1998) has studied the development of a low-level SW vortex which was involved its coupling with two upper-level disturbances. Both disturbance appeared later than and upstream of the low-level vortex. Faster eastward movements allowed them to catch up with the low-level vortex and led to a strong vertical coupling and deep tropopause folding. From the regional viewpoint, the topography of the Tibetan Plateau is extremely important. It is suggested that the formation of the SW vortex is a consequence of the blocking effects of the mesoscale mountain range, located at the southeastern corner of the Tibetan Plateau, on the southwesterly monsoonal flow. This is supported by the fact that the SW vortex develops mainly within the southwesterly monsoon current, that is strongest at the lower level during both its formation and mature stages, and that low level cyclonic vorticity is present throughout the period that the southwesterly monsoon flows around the Plateau. In addition, researchers have argued about the importance of cold air coming from the north and northeast.




daily hgt 850hPa 19990626
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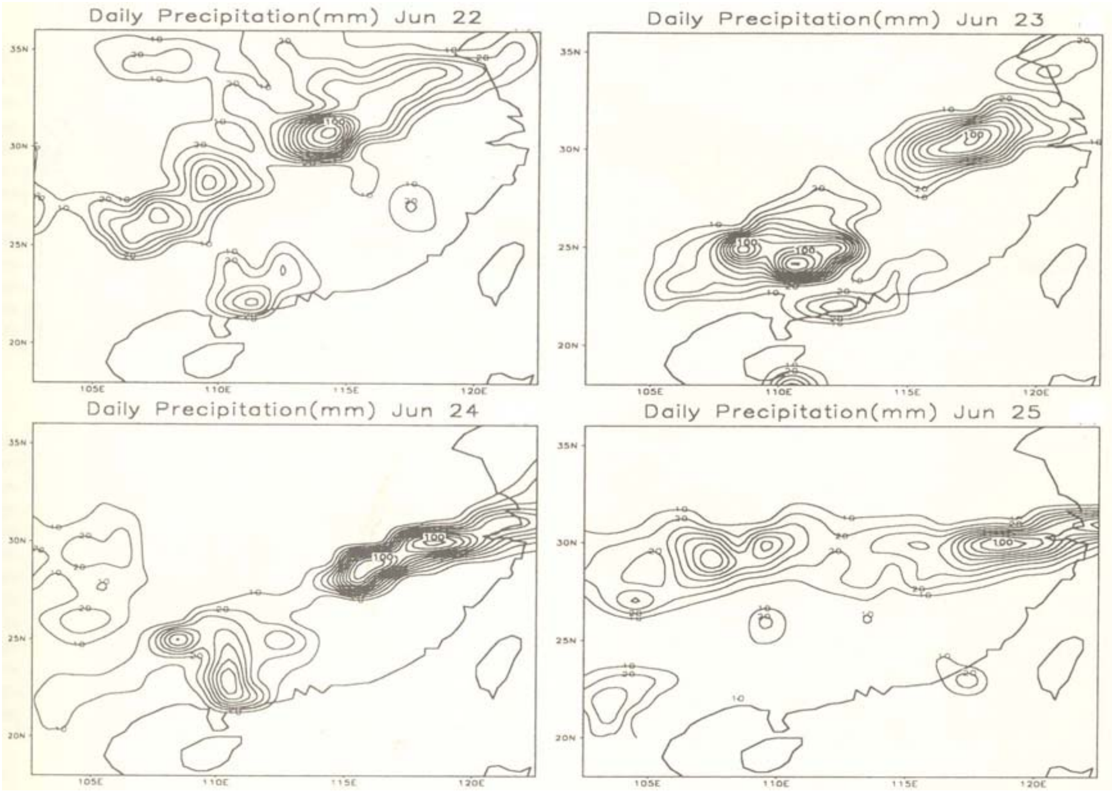
(a)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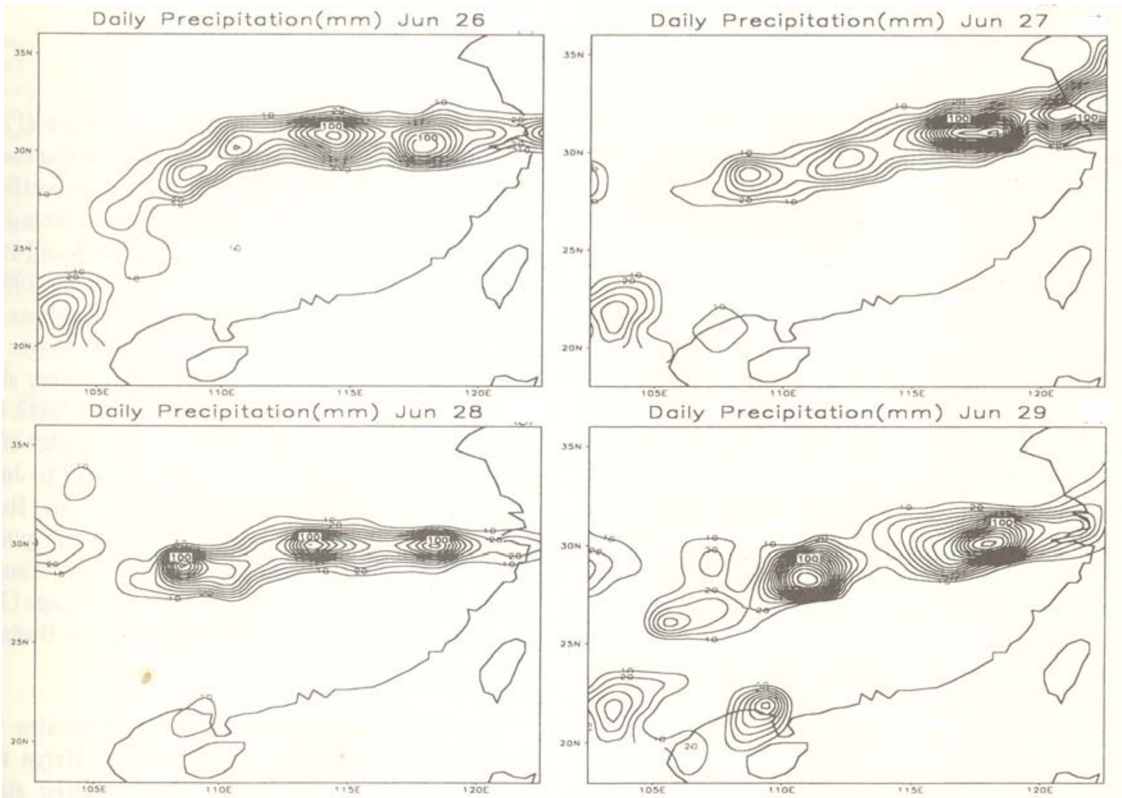
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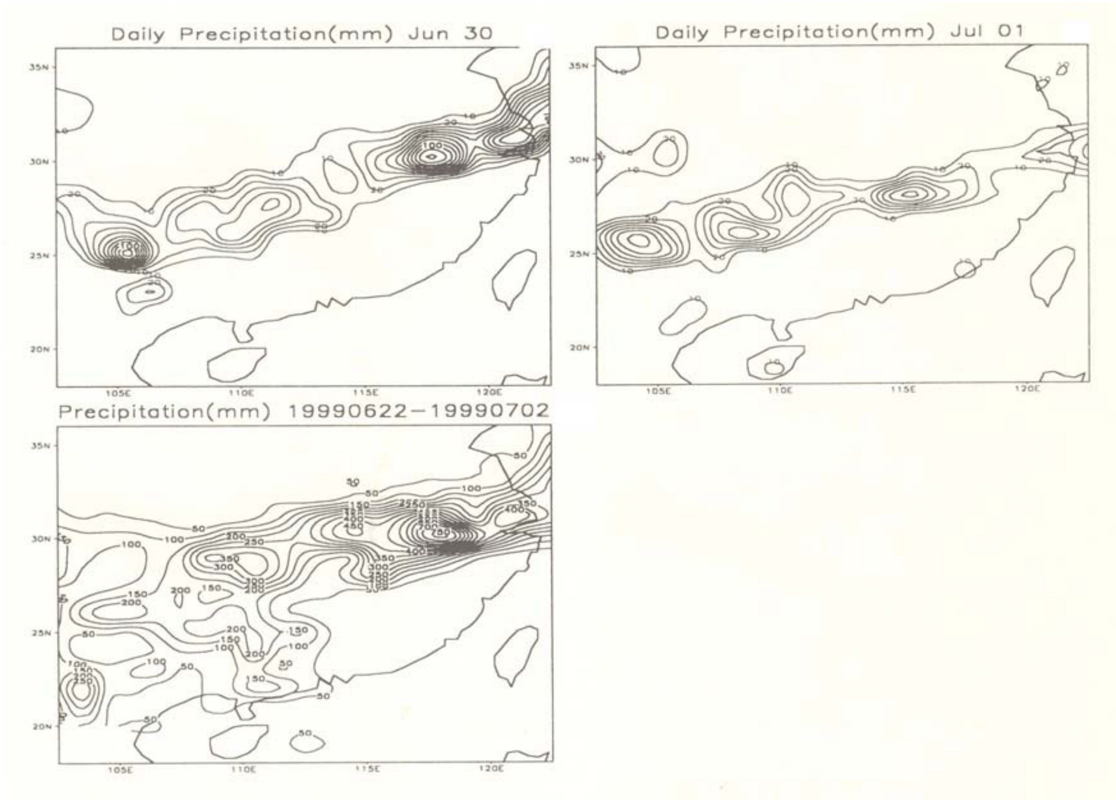


(b)

Figure 26. (a) Distributions of daily geopotential height and wind vector (unit: ms-1) at 850 hPa from June 22 to July 1, 1999; (b) same as (a), but for daily rainfall amounts (unit: mm). (Ding *et al.*, 2001)

The development of the SW vortex is expected to depend greatly on the effect of latent heat release, due to the fact that this vortex is usually accompanied by a large amount of rainfall and convective activity. In order to document better the effects of strong latent heat release associated with convection, Kuo *et al.* (1986) calculated mesoscale heat and moisture budget associated with a SW vortex which resulted in a flood catastrophe in the Sichuan Basin, on 11-15 July, 1981. With weak stability at the middle levels, latent heat release can induce strong, upward vertical motion, which in turn enhances low-level convergence spin-up and convective cloud development, establishing a positive feedback between the circulation of the SW vortex and the cumulus. Wang *et al.* (1993) further indicate that the mesoscale vortex in the lee of the Tibetan Plateau is driven diabatically.

4. Thermally Forced Circulation

Large plateau and mountain ranges generally constitute elevated heat or cold sources. The Tibetan Plateau, as the elevated heat sources, assumes significant thermal effects on large-scale circulation features and diurnal cycles (Yeh *et al.*, 1979; Luo and Yanai, 1984; He *et al.*, 1987). The interannual variation of heating over the Tibetan Plateau can exert an important effect on the inter-annual variability of circulation features over the Northern Hemisphere (Liu *et al.*, 2002). Over the Plateau region, the winter mean pressure pattern at low level shows that high pressure systems dominate. One may observe a closed high at 600 hPa which is basically separate from the Siberian cold high on the long-term mean chart. Therefore, they are two independent circulation systems. This fact reflects the prominent thermal impact of the Tibetan Plateau on the regional circulation system. In summer, there is a closed heat low over Tibet at low level. This heat low is also independent of the heat low or monsoon trough over India on the long-term mean 600 hPa chart. There are semi-permanent meso-scale features in the low pressure area whose impact is felt in the development of convective precipitating systems. The layer characterized by a monsoonal wind reversal (> 120 degrees between winter and summer) is rather distinct over the Tibetan Plateau, with the thickness of the monsoon layer greater than 4 km over the main body of the Plateau. The high elevation of the Tibetan Plateau obviously generates a great degree of baroclinicity with its surroundings, causing a vigorous circulation system to establish itself. The active layer of monsoonal flow may extend downstream far from the Plateau region (up to central China).

The upper level weather system over the Plateau in summer is predominantly the Tibetan high that is part of the huge South Asian high. The Tibetan high is of a thermal nature, with its peak intensity, horizontal extent, and high stability at 100 hPa. It has a very great effect on the circulation conditions in the Northern Hemisphere. In general, when the Tibetan high at 200 or 100 hPa superposes over the high at the lower and middle level (500 hPa), the Plateau would have a dry season or spell, whereas there would often be a rainy season or spell when the upper-level Tibetan high lies over a low-level low. The Tibetan high is the major center of action in the Northern summer and its formation and maintenance is to a great extent related to the effect of heat sources over Tibet and its surrounding areas. From both climatological mean and case study (Ding *et al.*, 1988), it has been found that while the Tibetan high advances northward, there is a center of high moving over Iran-Afghanistan and the western part of Tibet from the Arabian Peninsula and the latter has a greater intensity and a farther northern position than the eastern one. On the long-term 200 hPa flow charts in summer, there exist actually two huge highs over Asia, with the high center at 85-95^{o}E being the stronger. Recently, Zhang *et al.* (2002) have also indicated the bimodality of the 100 hPa South Asian high. This result, on the one hand, indicates the fact that the Tibetan high should be viewed as one cell out of the upper tropospheric belt of anticyclonic circulation. This cell is regionally intensified due to the thermal effect of the Plateau. On the other hand, the thermal effect of the Tibetan Plateau on the formation of the Tibetan high should not be overemphasized. At present, it is not clear what causes the formation of the high center over the western part of the Iranian Plateau.

The meridional circulation over the Tibetan Plateau and its surrounding areas is closely associated with the characteristics of the distribution of the heat source and sink over the Plateau and continental region. In summer, the Tibetan Plateau is a huge heat source. The upward motion predominates in the atmosphere over the plateau. The summer vertical circulation induced by the Tibetan Plateau is clearly seen in the meridional cross-section (Figure 27a), which is significantly different from that for the winter mean condition. On the whole, a huge monsoon meridional cell exists, with the upward flow of its northern leg extending up to 40-50^{o}N. The descending motion is predominantly observed in the Southern Hemisphere. Also, it may be seen that there are two smaller meridional cells on the northern and southern slopes of the Plateau, respectively. The northern slope only reaches 300 hPa in the vertical extent while the southern one extends upward up to over 200 hPa. Note that these two meridional cells only show up at the longitudinal range of the Tibetan Plateau, indicating that they result from the heating effect of the Plateau. The meridional cross section along 92.5^{o}E (Figure 27b) shows that the intense ascending motion over the Plateau is accompanied by subsidence over the Tarim Basin (40-50^{o}N). There is also a narrow zone of weak ascent between the stronger ascending motion over the Plateau and the vigorous ascent over Bangladesh (~20^{o}N) which is associated with the southwesterly monsoon. These are similar to the distribution of the mean July vertical circulation along 90^{o}E presented by Yeh, Gao *et al.* (1979) and simulated by Kuo and Qian (1981). Kuo and Qian investigated thermal effect of the Tibetan Plateau on the diurnal variation of meteorological fields in July (1981). They indicated that the huge diurnal variations of temperature and wind fields are mainly caused by the Tibetan Plateau, with a large-scale vertical circulation induced which is quite similar to a huge mountain-valley circulation. In night, the Plateau and northern and southern slopes below 300 hPa are dominated by descending motion, while it is replaced by strong ascending motion.

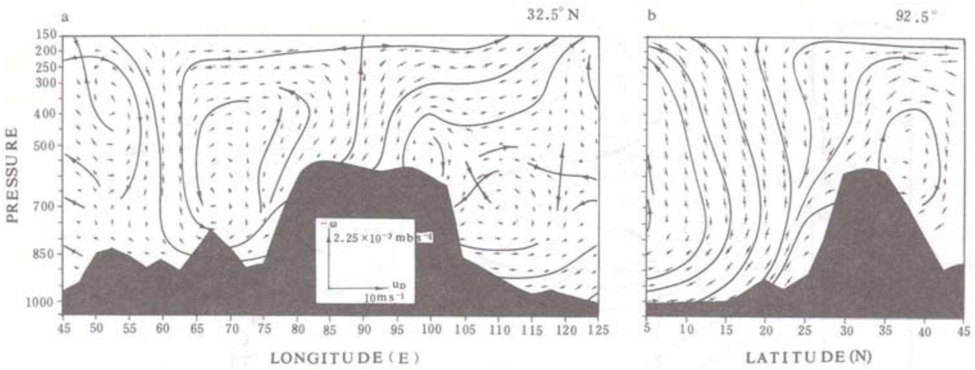


Figure 27. (a) The east-west vertical cross section showing the 80-day mean vectors (uD, ω) along 32.5°N (from 16 April to 4 July 1979), and (b) the north-south vertical cross-section showing the 80-day mean vectors (VD, ω) along 92.5°E. VD(uD, VD) is the divergent part of wind vectors (He, McGinnis, Song and Yanai, 1987)

5. Subtropical High over West-Pacific

The subtropical high is one important component of the East Asian monsoon system. To the western flank of this high, the southeasterly or southwesterly monsoon wind (the subtropical monsoon) often apparently develops in summer. The subtropical high over West-Pacific assumes a marked seasonal march in respect to its position which shows up as an alternation of gradual shift and abrupt jump. Three abrupt northward jumps may be observed, which are closely related to advance of the major rain belt in eastern China. The first northward jump occurs in the middle or last part of June, with the ridge line of the high moving over the region of 20-25°N. This shift indicates a northward jump of the rain belt over the Yangtze and Huaihe River Valleys, heralding the ending of the pre-summer rainy season in southern China. In the early and middle part of July, the ridge line of the high assumes second northward jump, reaching 25-30°N, starting the rainy season in the Yellow River and Huaihe River Valleys and ending in the rainy season in the Yangtze River Valley. The third northward jump occurs in the last part of July and early in August. This jump brings the ridge line of the subtropical high over the region around 30°N, thus leading to the beginning of the rainy season in northern China and the dry season in the Yangtze River Valley. At this time, the tropical easterlies or southeasterlies (southeasterly monsoon) prevail over an extensive region along the southern and western flanks of the subtropical high, and the ITCZ advances northward up to its seasonal northernmost latitude, 10-15°N over the western Pacific and 15-20°N over the South China Sea, where typhoons may frequently form. In this mid-summer season, the ridge line of the subtropical high may occasionally reach 35-40°N on the daily weather map. During the early part of September, the high retreats to the region to south of 25°N and then to south of 20°N in early October, starting the transition toward winter.

Three focuses has been recently placed on research on the subtropical high: (1) the role of seasonal change in the subtropical high in the onset of the Asian summer monsoon, (2) intraseasonal variation of the subtropical high and its effect on weather in East Asia, and (3) formation and maintenance mechanism of the subtropical high. Prior to the onset of the SCS summer monsoon, the breaking of the subtropical high over the region of Bay of Bengal is often observed (Mao *et al.*, 2003). This change in configuration of the subtropical high during the transition season is closely related to the development of the cyclonic circulation in this region which is due to enhancement of sensible heating over the Indian Peninsula and latent heat over Indochina Peninsula (Wen and He, 2002). The effect of convective heating in the monsoon cyclone over Bay of Bengal on the eastward retreat of the subtropical high has been shown (Liu *et al.*, 2002). The convective heating over the northern SCS and its interaction with the monsoon trough is also a condition for the eastward retreat of the subtropical high. The convective heating in above regions can lead to the change in the meridional temperature gradient from the normal condition to reverse sense (colder in south and warmer in north).

The zonal shift of the main body of the subtropical high has an important effect on weather in East Asia (Zhang and Tao, 2003). When the subtropical high has a further western position, an anomalous anticyclone at 850 hPa develops in the tropical region in East Asia while a anomalous cyclone develops in subtropical region, thus leading to intensification of ascending motion in Meiyu frontal areas. In such case, major precipitation processes occur in the Yangtze River basin. When the position of the subtropical high is farther to east, the major precipitation processes move down to the region to south of the Yangtze River. The zonal shift of the subtropical high can also have significant effect on summertime air temperature (Yang and Sun, 2003). They have a high positive correlation, i.e., relatively low air temperature corresponding to a further western position of the subtropical high. In addition, several investigators have studied the relationship between the occurrence of severe heavy rainfall/flood in the Huaihe River Basin during Meiyu season of 2003 and the characteristics of the subtropical high.

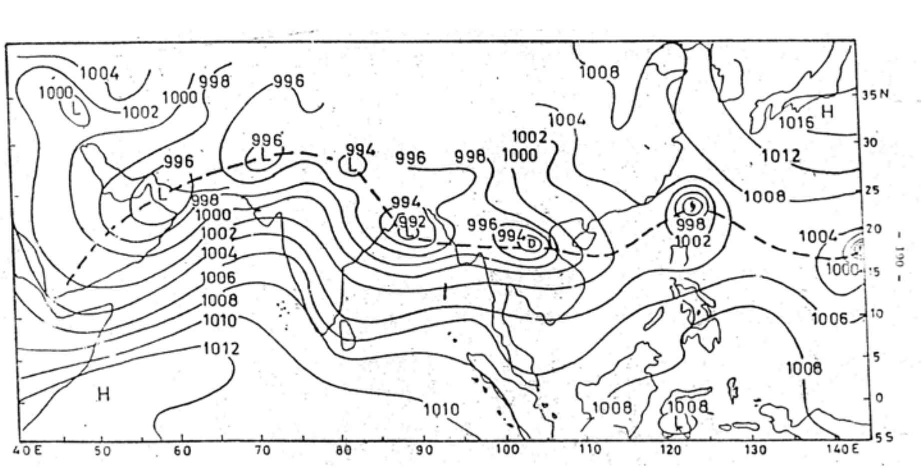
The mechanisms of formation and activity of the subtropical high is quite complicated. The diabatic heating is believed to be a critical factor (Wu and Liu, 2003; Liu *et al.*, 2004). The SST is also closely related to the activity of the subtropical high (Yan and Sun, 2003).

6. Monsoon Active and Break Processes

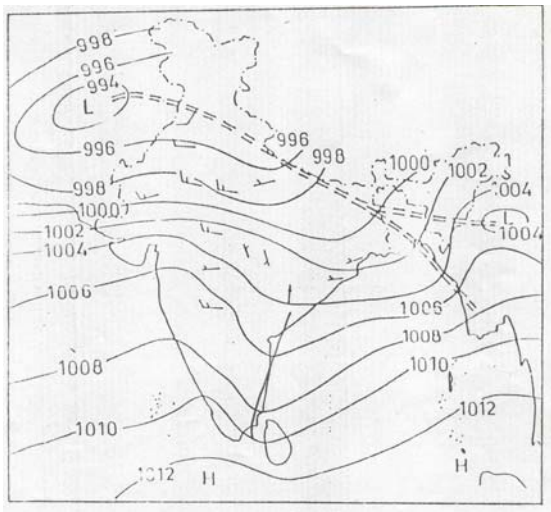
During the peak phase of the monsoon season in India, mid-season “active-break” cycle of the monsoon modulates the activity of monsoon rainfall and airflow. In this process, the monsoon trough and its attitudinal oscillation play an essential role. The northward propagating of intraseasonal oscillations (ISO) on 30-40 day scale provides the super-envelope for overlapping formation of monsoon disturbances (Sikka *et al.*, 1986) during the peak phase of monsoon activity. The position and activity of the monsoon trough over the Indo-Gangetic plans, in itself, fluctuates on synoptic and intr-seasonal scales, thus controlling cyclogenesis within it and hence the statistical rainfall distribution. When the major rain belt is situated over central India, the condition is called one of ‘active’ monsoon and when it is highly suppressed over the same region it is called the ‘break’ monsoon. Figure 28 shows the schematic of vertical motion along the monsoon trough across the Gangetic Plains during ‘active’ and ‘break’ monsoon conditions.

When the monsoon trough hugs the Himalayan rim, the low level easterlies disappear entirely along the Indo-Gangetic Plains and are replaced by the W to NW flow along the periphery of the Himalayas. Such a situation, if it last for 2 or 3 days bringing back the monsoon trough in the Gangetic Plains, is known as the temporary oscillation of the monsoon trough or temporary ‘break’ in the monsoon. If the sub-montane position of the trough sticks for more than 3 to 4 days, it leads to a prolonged ‘break monsoon’ situation in which the rainfall almost ceases over most of central India, increases at first over the foothill regions and subsequently under prolonged or persistent ‘break’ (duration more than 5 days), it even ceases across the foothill region too (Sikka, 2004, personal communication).

Under ‘break’ monsoon conditions, it does rain heavily for about three days along the foothills of Himalayas as most of the moisture convergence occurs in that belt and the forced ascent by orography further augments the rainfall. As the ‘break’ monsoon persists longer and reaches its peak, rainfall ceases even across the foothill region of the Himalayas and an organized cloud band is observed to move steadily northward (1°lat/day) from near-equatorial oceanic belt of NHET (5-10°N). In a couple of days the northward moving band approaches the North Bay of Bengal. With the southward shift of the sub-Himalayan cloud band, the monsoon cloud band is revived along its near-normal position. Such a sequence of events is the well known 30 to 40-day ISO of the monsoon. This is an important singularity which first occurs in the season at the onset of the monsoon and subsequently 2 or 3 times within a season. In different ways it controls the performance of the seasonal monsoon.



(a)



(b)

Figure 28. (a) A typical sea-level pressure pattern in which there are five weather disturbances during the active phase of the monsoon. (b) Typical sea-level isobaric pattern over South Asia during a 'break' monsoon situation with the monsoon trough lying across the foothills of Himalayas and a north-south trough along the east coast of India. (Taken from Ding and Sikka, 2004)

The frequency and total duration of 'break' monsoon episodes definitely control the sub-normal performance of monsoon on intra-seasonal and inter-annual scales. The 'break' monsoon phenomenon is reverse of the 'active' monsoon spell over central and northern India and the two demonstrate the two modes of the South Asian monsoon system. Goswami and Mohan (2001), Krishnamurthy and Shukla (2000), and Lawrence and Webster (2001) have found similarities in the ISO and the inter-annual oscillation of the monsoon system. This is not surprising as it is due to the dominance of the two modes ('active' and 'break') of monsoon activity in the monsoon season. The excess and drought monsoon seasons on inter-annual basis are in some way replicate 'active' and 'break' monsoon events on intra-seasonal scale. Whether this dominance is pre-controlled by the land-surface oceanic boundary forcing or results due to internal dynamics of the monsoon and air-sea interactions remains still a moot question. The revival of the monsoon after a short 'break' (3 to 4 days) is due to the synoptic scale oscillatory character of the monsoon trough. However, the revival after a prolonged 'break' is distinctly linked with the northward propagating ISO (Sikka and Gadgil, 1980). Koteswaram (1950) had alluded to the westward movement of weak low pressure systems along 10-15^{0}N over the Bay of Bengal as a precursor to the revival of monsoon after a prolonged 'break'. In some years the 10-20 day westward moving ISO from the South China Sea (Krishnamurti and Ardanuy, 1980; Lau and Chan, 1986) is also responsible for restoring the monsoon trough at near-normal position and hence the revival of the monsoon after a moderate duration 'break' episode.

In East Asia, major monsoon rainy seasons such as the presummer rainy seaon in South China, and Meiyu/Baiu rainy season in the Yangtze River Basin and Japan occurs during May-mid-July. Afterwards, a break of the monsoonal rainy period occurs from late July to early August in Japan (Chen *et al.*, 2003). This break of different spans is also observed in South China, Central China, Northeast China, Taiwan, and Korea (Figure 29). After the break spell, monsoon rain resumes for a period from August to September-October. Therefore, the monsoon rainfall variation during the warm season in East Asia is generally characterized by two active rainfall periods separated by a break spell. It is clearly seen from Figure 29 that the Meiyu rain band, forming in early May, progresses northward until the end of July, and diminishes between 40° and 45°N in Northeast China and Korea, and about 40°N in Japan. The passage of the Meiyu rain band is followed by a break spell (monsoon break) which also propagates northward. Then, the monsoon rainfall revival after the break is clearly observed. Chen *et al.* (2003) has shown that the monsoon revival in East Asia is caused by a different mechanism associated with the development of other monsoon circulation components including the ITCZ and weather systems in mid-latitudes. The Changma is a shorter monsoonal rainy season, with mean Changma period being 20 days to one month long. The break in late July is very short, with the duration of a half month. Starting from late August, the revival of the monsoon rainy period is also observed in Figure 29. The second rain spell is not long based on the study by Chen *et al.* (2003). But, Qian *et al.* (2002) pointed out that this precipitation surge can maintain until early September, forming the autumn rainy season in Korea.

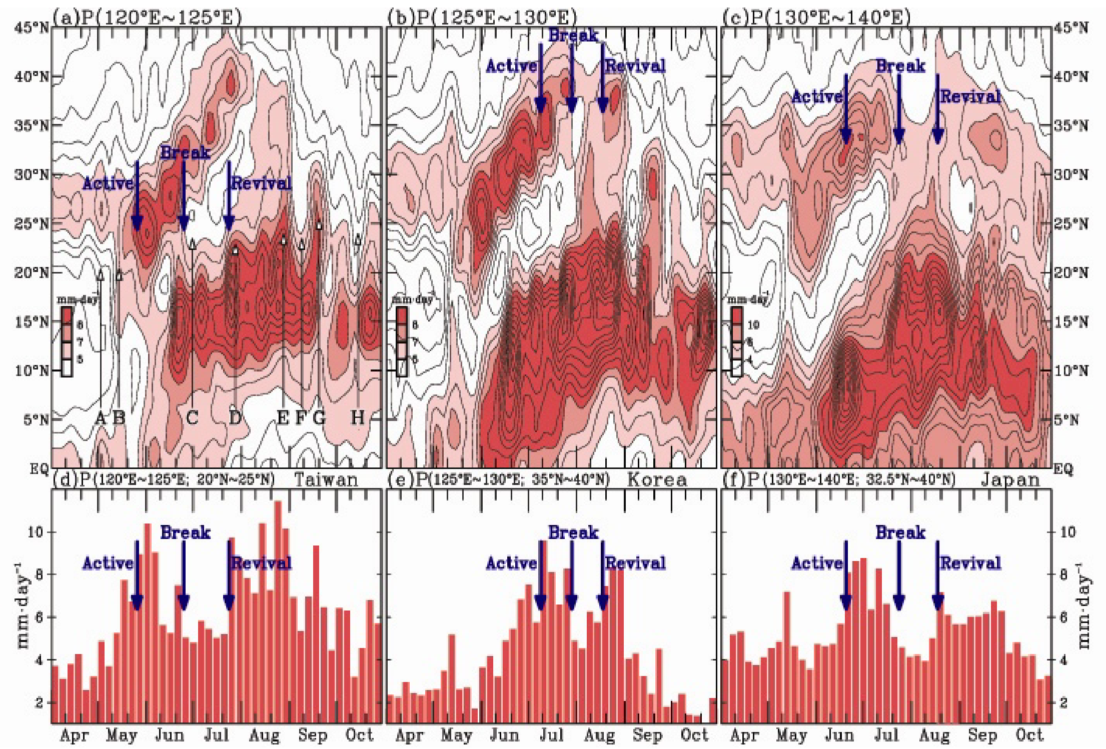


Figure 29. Latitudinal-time cross-sections of CMAP rainfall averaged over longitudinal zones of (a) 120°-125°E, (b) 125°-130°E, and (c) 130°-140°E, and rainfall histograms of three regions: (d) Taiwan (120°-125°E, 20-25°N), (e) Korea (125°-130°E, 35°-40°N), and (f) Japan (130°-140°E, 32.5°-40°N). Different phases of summer monsoons in three regions are indicated by active, break and revival. The contour interval of CMAP rainfall in (a)-(c) is 1 mm day-1, while rainfall amounts larger than 5 mm day-1 are stippled by different colors indicated by the scale shown in the lower left corner of the three upper panels. (Chen *et al.*, 2003)

7. Tropical Easterly Jet Stream (TEJ)

The tropical easterly jet steam (TEJ) in the upper troposphere is one of the most important components of summer monsoon systems in the Indian and East Asian monsoon regions, and also one of the major circulation features in the tropical upper troposphere in the northern summer. Its establishment and activity are closely associated with the seasonal change in the upper troposphere in the Northern Hemisphere, and summer monsoon and precipitation events in South Asia and Southeast Asia. Numerous investigators have studied the characteristics of activity and the cause of the formation of the TEJ as well as its close relationship to the precipitation in monsoon regions (Koteswaram, 1958; Flohn, 1964; Krishnamurti, 1978). The TEJ has its maximum intensity in the layer of 150-100 hPa. Therefore, it is preferable to use the synoptic maps at 100 or 150 hPa to study the behavior of the TEJ, but the 200 hPa synoptic map has been sometimes utilized due to the sparse data sources at higher levels.

The TEJ ahs a very high steadiness, usually attaining 96-100%. The interannual variability of the position of the jet core is rather small. Therefore, the TEJ is an almost steady and persistent circulation feature during the summer.

Based on the upper-level data for 1966-1975, Zheng and Guo (1982) plotted the long-term mean maps of the TEJ. In June (not shown), the eastern part of the TEJ axis is located at about 15°N and the western part at about 10^{o}N, thus causing the whole axis of the TEJ to be oriented from east-northeast to west-southwest. In July (Figure 30), the axis of the TEJ moves northward by about 5 latitudes, with the eastern part situated at 18-20^{o}N and the western part at 15°N. In August, the TEJ returns back to its June position. The greatest intensity of the TEJ is observed in July, with a jet core of 35ms^{-1} at 15°N，80°E. The area with strong winds greater than 20ms^{-1} zonally extends by 100 longitudes (15-125°E).

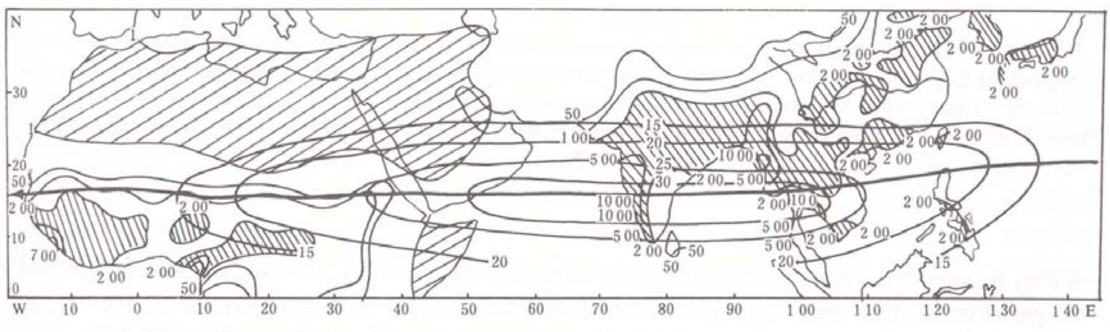


Figure 30. The long-term mean TEJ for July 1966-1975. The bold black arrow represents the axis of the TEJ; the heavy full lines denote isotachs (Units: ms^{-1}); the full lines isohyets (Units: mm). Shadded area denotes the regions with rainfall amount greater than 200mm (Zheng and Guo, 1982).

Yang (1982) has studied the splitting phenomenon of the TEJ at 100 hPa and indicated that it has two branches over the region from South Asia and Southeast Asia to the tropical western Pacific with the northern branch located over the latitude of 10-20°N at 100 hPa, and the southern branch found near 8°N at 150 hPa, slightly lower than the northern one. Xiao and Xie (1983) analysed the structure of the TEJ particularly in connection with the condition of the two branches of the jet in Southeast Asia, which is an important characteristic feature for the East Asian summer monsoon system. The northern branch of the jet located at about 17°N at 100 hPa, with maximum wind speed greater than 20 ms^{-1}. Therefore, the northern leg of the TEJ is higher and stronger. A very shallow layer of westerly wind lies below the layer of the easterlies in the northern of the TEJ, whereas there is a deeper layer of stronger westerly wind below 500 hPa for the southern leg of the TEJ.

It is worthwhile showing the mean meridional cell nearly normal to the axis of the TEJ. In the northern part of this cell there is the very strong upward motion, while in the southern part, the air descends, thus producing a marked negative meridional cell. The sense of this kind of the meridional cell is in good agreement with the result obtained by Flohn (1964) for the condition of the entrance region of the TEJ (90-130°E). He indicates that the meridional cell causes an extensive ascending motion and convective activity over Southeast Asia and the South China Sea. The tropical disturbances at low-level develop while they are entering the northern sector of the entrance region of the southern branch of the TEJ. This situation is very simlar to that for the westerly jet-stream at the middle latitude.

The TEJ is one of the most important circulation features that affect the monsoon activity and precipitation in the tropical African and Asian regions. Over the Indian monsoon region, on the one hand, the active and break cycles of the summer monsoon, the shift of the monsoon rain belt, and the periodic fluctuation of the intensity of monsoonal airflows are all related to the position and change in the intensity of the TEJ (Koteswaram, 1958). On the other hand, the activity of the TEJ is also related to the shift of rain belts in China. In May when the TEJ begins to establish itself over the southern part of the Indo-China Peninsula, it may have an effect on rainfalls during the pre-summer rainy season in southern China. Its activity may also affect the onset of Meiyu over the Yangtze River Valley and the rainy season in China. Koteswaram (1958) studied the relationship between the TEJ and the summer precipitation over the African-Asian monsoon regions on the basis of particular years. The precipitation patterns are different for the different quadrants of the TEJ. Roughly with 60°E taken as the demarcation line between the entrance and exit regions of the TEJ, much precipitations are mainly located in the left quadrant of the exit region (to the west of 60°E) and the right quadrant of the entrance region of the TEJ (to the east of 60°E). This result is in coincidence with the patterns of the transverse secondary circulation across the TEJ. The major rainfalls occur in the region of the ascending motion from the equator to 30°N.

The African easterly jet is also an important component of upper-tropospheric system and the African monsoon system. Its instability is closely related to genesis and development of low-level tropical disturbances that can propagate westward across the tropical Atlantic, as the African tropical waves. The typical precipitation patterns are associated with variation and location of the African easterly jet (Also see Figure 30).

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